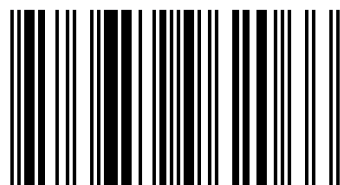


This book offers the basic knowledge to understand the genesis of different mineral ores. It is written in a simple way that the non-specialist can go through smoothly. I trust that this book will be useful to both scholars and practitioners, wherever they work. This book is written based on geoscientific aspects of mineral deposits, including their origin and geological characteristics. I concentrated mainly on the different geologic processes by which the ores are generated and concentrated. Tens of images from different sources are found to let the reader easily imagine the multiple formation processes of the different ores. It can be considered as a main course for the underground level, however, it will add basic important information to the postgraduate candidates. This book provides a panoramic view over the whole field of economic geology.



AbdelMonem Soltan is a professor of Applied Mineralogy at Ain Shams University, Cairo, Egypt. He got his Ph.D. in applied mineralogy from Ain Shams University, Egypt, in 2007. He published 20 research articles in peer-reviewed journals and joined 16 international conferences. AbdelMonem Soltan is currently doing research on ceramics and concretes.



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Moursi Soltan

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Economic Geology: Lecture Notes

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Economic Geology: Lecture Notes



Abdel Monem Soltan

Economic Geology: Lecture Notes

Abdel Monem Soltan

(PhD)

Geology Department, Faculty of Science, Ain Shams University, Cairo,
11566, Egypt.

To Samah, Ayten and Omnia

Preface

Economic geologist always aims at understanding the concepts that would help in discovering new mineral deposits to exploit them to the best. Actually, Economic Geology is a fascinating subject in its own right even for geologists who do not intend to be professional economic geologists.

The "Economic Geology: Lecture Notes" is a "capstone" book that aims to illustrate and integrate among many other geological disciplines (mineralogy; stratigraphy; sedimentology; structural geology; tectonics; petrology; geochemistry; ...etc). This book can be considered as an introductory course in Economic Geology. The main target audience is under- and post-graduate geologists and engineers. However, the book also serves as a useful information resource for professional economic geologists.

This book describes the different geological processes causing the accumulation of mineral deposits. As well, it provides an overview of the principal features of the different mineral deposits. It includes: basic geologic concepts (Chapter 1); introduction (Chapter 2); magmatic ore deposits (Chapter 3); supergene ore deposits (Chapter 4); sedimentary ore deposits (Chapter 5); diagenetic ore deposits (Chapter 6); metamorphic ore deposits (Chapter 7) and metallogeny and plate tectonics (Chapter 8). The geochemical aspects of the different mineral deposits are beyond the interest of this book. The book contains 250 figures and 15 tables.

The book material is mainly compiled with modifications from:

- i. Pohl W. L., 2011. "Economic Geology Principles and Practice", Blackwell Publishing Ltd, p. 663.
- ii. Misra K. C., 2000. "Understanding Mineral Deposits", Springer, p. 845.
- iii. Harald G. D., 2010. "The "chessboard" classification scheme of mineral deposits: Mineralogy and geology from aluminum to zirconium", Earth-Science Reviews, Volume 100, Issues 1-4, p. 420.
- iv. Different internet resources.

I have a debt of gratitude with many colleagues, too many to be mentioned here. But in particular I would like to acknowledge Prof. Dr. Mohamed Abu-Zeid; Prof. Dr. Baher El-Kaliouby; Prof. Dr. Ali Farrag and Dr. Zeinab Taman, Geology Department, Faculty of Science, Ain Shams University, 11566, Cairo, Egypt.

Abdel Monem Soltan

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Chapter 1: Basic Geologic Concepts

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1- Dynamic structure of the Earth

The Earth is an irregular sphere, with a radius that varies between 6,356 and 6,378 km. This solid sphere is chemically divided into layers that become less dense from the centre towards the surface.

The three main layers are:

- (i) the core (which comprises an Inner Core and an Outer Core);
- (ii) the mantle, and
- (iii) the crust.

Each layer has a distinctive chemical composition, and a different density (Fig. 1).

The core is primarily composed of the heavy elements iron and nickel. The outer core is made of molten iron, which produces the Earth's magnetic field.

The mantle is less-dense than the core. The mantle extends to a depth of about 2,900 km (Fig. 1).

The mantle is rich in iron- and magnesium bearing silicate minerals. The outer layer of the Earth is termed the crust, which is divided into oceanic crust and continental crust (Fig. 2). Overall, continental crust is richer in the element silica, and is less dense, than oceanic crust. Oceanic crust (about 10 km thick) is composed of iron-, magnesium-, calcium-, and aluminium-rich silicate minerals that typically form a dark colored, heavy rock called basalt. Continental crust (about 20 - 60 km

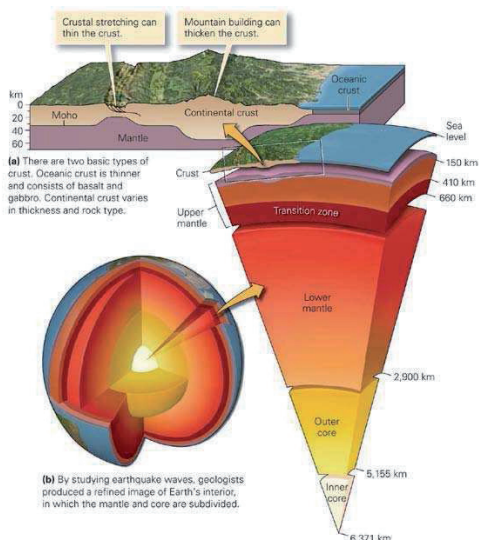


Fig. 1. The structure of the Earth

- Oceanic Crust
 - primarily basalt
 - 4-7 km thickness (thin relative to continental crust)
 - denser (heavier) than continental crust
- Continental Crust
 - primarily granite
 - 20-70 km thickness
 - less dense (will not undergo subduction)

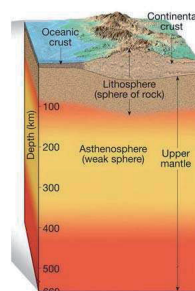


Fig. 2. Oceanic and continental crust

thick) is composed of potassium-, sodium-, and aluminium-rich silicate minerals that form a diverse range of rock types such as granite.

The crust and upper part of the mantle of the Earth is further subdivided into the lithosphere and the asthenosphere (Fig. 3). The lithosphere is a strong layer, extending to a depth of 100 to 150 km, that comprises the crust and part of the upper mantle (the upper rigid part). The lithosphere is separated into seven large plates (Fig. 4), and several smaller plates. These plates, which terminate at different types of plate boundary, move over the underlying asthenosphere.

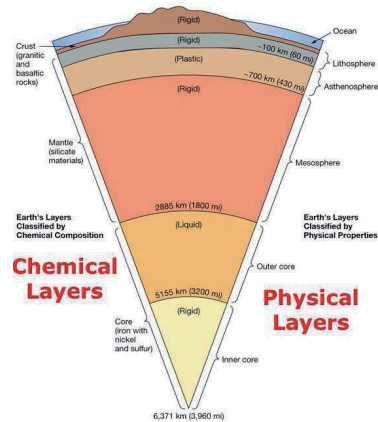


Fig. 3. Physical and chemical layers of the Earth.

The asthenosphere (the middle part of the mantle - plastic, i.e., semi-liquid and ductile) is a weaker layer, upon which the lithospheric plates move, and from which magmas that form the oceanic crust are derived.

Heat from the Earth's core creates circulation patterns (i.e., convection currents) in the mantle drive the motions of the overlying plates. The slow movement of the lithospheric plates over the mobile asthenosphere is known as plate tectonics, a process that maintains

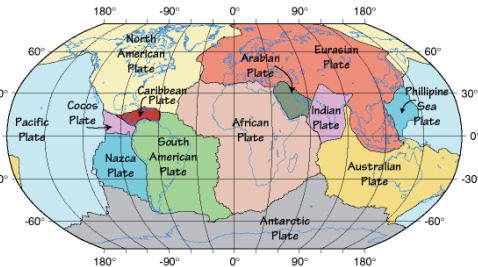


Fig. 4. Tectonic plates of the world.

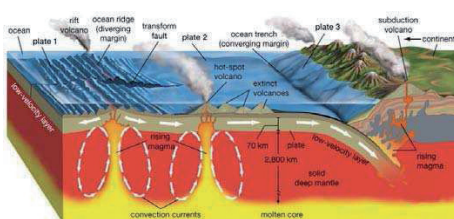


Fig. 5. Because ocean plates are denser than continental plates, when these two types of plates converge, the ocean plates are subducted beneath the continental plates. Subduction zones and trenches are convergent margins. The collision of plates is often accompanied by earthquakes and volcanoes.

through a material with any bulk motion of its particles. Convection is common in fluids).

Convection currents in the asthenosphere transfer heat to the surface, where plumes of less dense magma break apart the plates at the spreading centers (Fig. 5). This creates divergent plate boundaries. As the plates move away

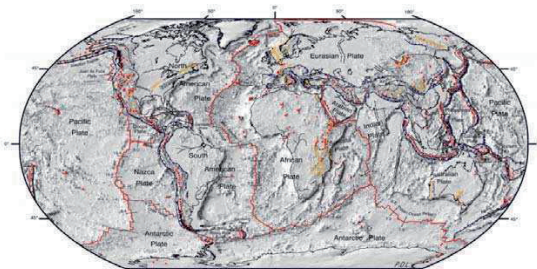


Fig. 6. Earth plates.

from the spreading centers, they cool, and the higher density basalt rocks that make up ocean crust get consumed at the ocean trenches/subduction zones. The crust is recycled back into the asthenosphere.

2- Plate tectonics

Plate tectonics (previously known as continental drift) originated from the geographical observation that the coastal profiles of South America and Africa seem to fit one another. First proposed by Alfred Wegener in the 1920s, the crust was imagined to be made up of continent-sized slabs that "float" on a liquid layer and thus "drift" around (Fig. 6). Plate tectonics, appeared in the 1960s when the mid-Atlantic ridge was discovered, along with compelling evidence for injection rock caused "spreading" leaving parallel north-south trending stripes of injected rock, the youngest of which was adjacent to the injection ridge and the oldest farthest from it. The plate tectonics solution to the seafloor spreading dilemma was the proposition that new crustal mass created by injection must be compensated by "subduction", the diving of ocean crust (more dense) under opposing continental plates (less dense). Subduction zones and trenches are convergent margins. The collision of plates is often accompanied by earthquakes and volcanoes (Fig. 7).

• Plate motions

1. There are two basic types of lithosphere: Continental lithosphere has a low density because it is made of relatively light-weight minerals. Oceanic lithosphere is denser because it is composed of heavier

- minerals. A plate may be made up entirely of oceanic or continental lithosphere, but most are partly oceanic and partly continental (Fig. 7).
2. Beneath the lithospheric plates lies the asthenosphere, a layer of the mantle composed of denser semi-solid rock. Because the plates are less dense than the asthenosphere beneath them, they are floating on top of the asthenosphere (Fig. 7).
 3. Deep within the asthenosphere the pressure and temperature are so

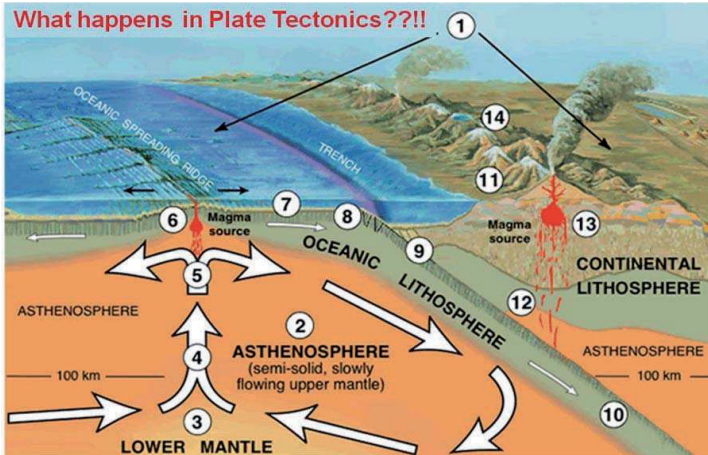


Fig. 7. This diagram shows the interaction between continental and oceanic plates, the processes illustrated generally apply for the interaction between two oceanic plates.

- high that the rock can soften and partly melt. The softened dense rock can flow very slowly. Where temperature instabilities exist near the core/mantle boundary, slowly moving convection currents may form within the semi-solid asthenosphere (Fig. 7).
4. Once formed, convection currents bring hot material from deeper within the mantle up toward the surface (Fig. 7).
 5. As they rise and approach the surface, convection currents diverge at the base of the lithosphere. The diverging currents exert a weak tension or "pull" on the solid plate above it. Tension and high heat flow weakens the floating, solid plate, causing it to break apart. The two sides of the now-split plate then move away from each other, forming a divergent plate boundary (Fig. 7).

6. The space between these diverging plates is filled with molten rocks (magma) from below. Contact with seawater cools the magma, which quickly solidifies, forming new oceanic lithosphere. This continuous process, operating over millions of years, builds a chain of submarine volcanoes and rift valleys called a mid-ocean ridge or an ocean spreading ridge (Fig. 7).
7. As new molten rock continues to be extruded at the mid-ocean ridge and added to the oceanic plate (6), the older (earlier formed) part of the plate moves away from the ridge (Fig. 7).
8. As the oceanic plate moves farther and farther away from the active, hot spreading ridge, it gradually cools down. The colder the plate gets, the denser ("heavier") it becomes. Eventually, the edge of the plate that is farthest from the spreading ridges cools so much that it becomes denser than the asthenosphere beneath it (Fig. 7).
9. As it is known, denser materials sink, and that's exactly what happens to the oceanic plate—it starts to sink into the asthenosphere! Where one plate sinks beneath another a subduction zone forms (Fig. 7).
10. The sinking lead edge of the oceanic plate actually "pulls" the rest of the plate behind it—evidence suggests this is the main driving force of subduction. It is not sure how deep the oceanic plate sinks before it begins to melt and lose its identity as a rigid slab, but it remains solid far beyond depths of 100 km beneath the Earth's surface (Fig. 7).
11. Subduction zones are one type of convergent plate boundary, the type of plate boundary that forms where two plates are moving toward one another. Notice that although the cool oceanic plate is sinking, the cool but less dense continental plate floats like a cork on top of the denser asthenosphere (Fig. 7).
12. When the subducting oceanic plate sinks deep below the Earth's surface, the great temperature and pressure at depth cause the fluids to "sweat" from the sinking plate. The fluids sweated out percolate upward, helping to locally melt the overlying solid mantle above the subducting plate to form pockets of liquid rock (magma) (Fig. 7).
13. The generated magma is less dense than the surrounding rock, so it rises toward the surface. Most of the magma cools and solidifies as large bodies of plutonic (intrusive) rocks far below the Earth's surface (Fig. 7).

14. Some of the molten rock may reach the Earth's surface to erupt as the pent-up gas pressure in the magma is suddenly released, forming volcanic (extrusive) rocks (Fig. 7).

• **Types of Plate Boundaries**

There are three types of plate boundary: convergent, divergent, and transform plate boundaries. Divergent plate boundaries occur where two lithospheric plates move away from each other, driven by magma rising from deep within the mantle (Fig. 8). Volcanic activity at a divergent plate boundary creates new lithosphere along what is known as a spreading ridge. Convergent plate boundaries occur where two lithospheric plates move towards each other, with one plate overriding the other (Fig. 9). The overridden plate (sinking plate) is driven back into the mantle, and is subsequently destroyed along what is known as a subduction zone. During this process, earthquakes and volcanic activity are generated in the overriding plate.

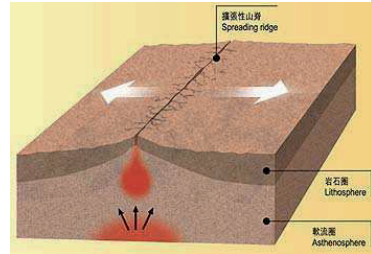


Fig. 8. Divergent plate boundaries.

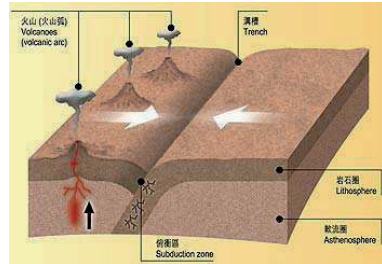


Fig. 9. Convergent plate boundaries.

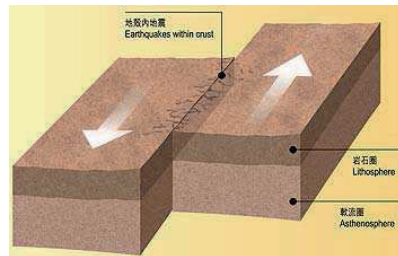


Fig. 10. Transform plate boundaries.

Transform plate boundaries occur where two lithospheric plates slide laterally past each other (Fig. 10). Earthquakes are generated along this type of plate boundary. Importantly, lithosphere is preserved along transform boundaries, it is not created or destroyed as it is at divergent and convergent plate



Fig. 11. Magma.

boundaries.

3- Magma

Magma is hot molten rock within the earth. It can well-up from deep to extrude from fractures as lava flows and/or pyroclastic ejecta (Fig. 11). The source for magma is not the earth's liquid outer core, a common misconception; instead, magma is generated at the relatively shallow depths of 100 to 300 km, through the partial melting of the earth's crust and mantle. It is most often formed by decompression-melting of asthenosphere associated with divergent plate boundaries or mantle plumes, or by partial-melting of water-rich crust and/or asthenospheric material in association with subduction at convergent plate boundaries.

The ingredients necessary for the production of magma involve the interplay between heat, pressure, intra-granular fluids (present as gases within very hot rock or magma) and the composition of the material subject to melting:

1. Heating obviously brings solids closer to their melting points, the more heat, the more likely a solid will melt (Fig. 12).
2. In general, higher pressures prevent melting because the constituent atoms of minerals in rocks are squeezed together and remain solids under high pressure (Fig. 12). Consequently, lowering pressure on hot rock induces melting.

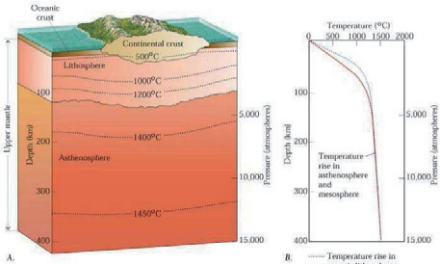


Fig. 12. Geothermal gradient of the Earth.

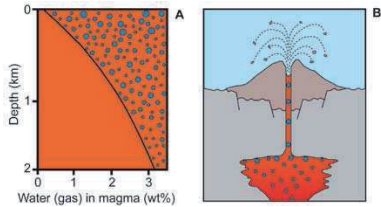


Fig. 13. Gases in the magma.

Table 1: Chemical composition of magma.
Chemical Analyses of Magmas

Oxide %	Basalt	Andesite	Dacite	Rhyolite
SiO ₂	50.10wt.%	59.20	67.20	75.20
TiO ₂	1.00	0.70	0.50	0.20
Al ₂ O ₃	17.10	17.10	16.20	13.50
Fe ₂ O ₃	3.40	2.90	2.00	1.00
FeO	7.00	4.20	1.80	1.10
MgO	7.10	3.70	1.50	0.50
CaO	10.60	7.10	3.80	1.60
Na ₂ O	2.50	3.20	4.30	4.20
K ₂ O	0.80	1.30	2.10	2.70
P ₂ O ₅	0.20	0.20	0.20	0.10

←—————→
continuum of compositions

3. Intra-granular fluids (gases within very hot rock or magma) lower the melting point of solids, so the presence of fluids (gases), generally water, allows solid rock to melt at a lower temperature (or heat content) than it otherwise would (Fig. 13).
4. Finally, there are two general trends to explore in relation to rock composition: rock that contains a relative abundance of silica (SiO_2) and aluminum (aluminum oxide) will melt at a lower temperature (heat content); while a rock containing a relative abundance of ferromagnesian (Fe, Mg, and Ca) ions will melt at higher temperatures (heat content).

The melting of continental crust generates felsic magma enriched in silica and aluminum, while melting of mantle rock (asthenosphere) and oceanic crust forms ferromagnesian-rich, mafic magma (Table 1). The earth's crust naturally contains a higher water content (because of its proximity to the hydrosphere) than the mantle, accounting for higher water (and thus gas) content in felsic to intermediate magmas. The relatively high content of silica and water in continental crust also correlates with the lower melting temperatures of felsic to intermediate magmas (Table 1). Mantle material melts at greater depth and higher temperatures and pressures, not requiring as much "assistance" from silica and water in the melting process.

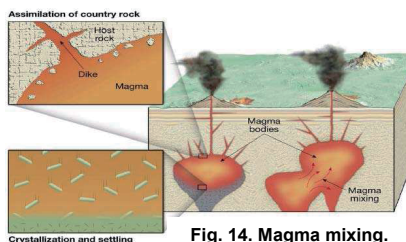


Fig. 14. Magma mixing.

• Magma types

The composition of magma (and extruded lava) depends on three main factors:

- 1) the degree of partial melting of the crust or mantle;
- 2) the degree of magma mixing (Fig. 14);
- 3) magmatic differentiation by fractional crystallization.

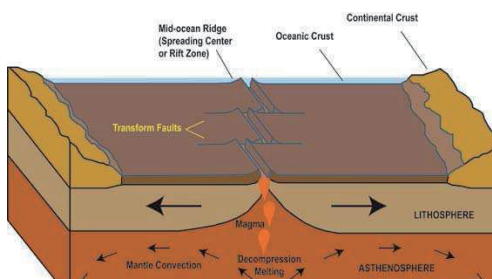


Fig. 15. Mafic magma (at divergent plate boundaries).

There are three main types of magma:

1. Mafic magmas are generated by decompression-melting of highly mafic asthenosphere and assimilation-melting of mafic oceanic lithosphere and crust in association with divergent plate boundaries and some mantle plumes (Fig. 15). The magma source is naturally low in water content, however, these magmas have a much easier time of it; greater heat and less silica allows it to readily reach the surface as volcanic eruptions (despite its lack of gases). Mafic magmas have lower viscosities because of their greater heat content and lack of silica (they have a greater abundance of iron and magnesium ions).

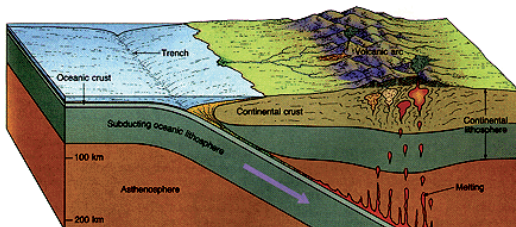


Fig. 16. Felsic magma (at convergent plate boundaries).

2. Felsic magmas have higher viscosities because of their lower heat content and enrichment with respect to silica. Felsic magmas are generated by the partial melting

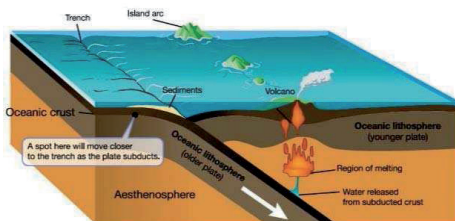


Fig. 17. Oceanic-oceanic plate collision.

of the more siliceous upper portion of water-saturated oceanic crust (more siliceous because of the thick sedimentary cover it carries) where it is subducted at convergent plate boundaries and by assimilation-melting of siliceous, water-rich, continental crust into the magma derived from partial melting of mafic oceanic crust and asthenosphere as it rises toward the surface (Fig. 16).

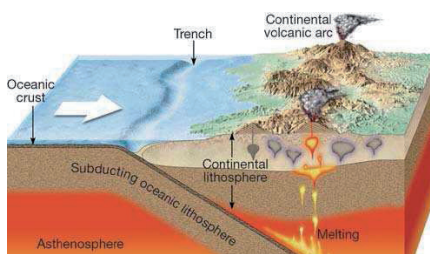


Fig. 18. Oceanic-continental plate collision.

- Intermediate magma: During oceanic-oceanic plate collisions, a basic magma rises through the overlying oceanic plate and is little changed by assimilation-melting (the original mafic magma simply assimilates more mafic material on its way upward) and volcanic eruptions on the sea floor form island chains called island arcs (Fig. 17). Volcanism is initially mafic in composition, but as time progresses and the volcanic arc ages and is subject to erosion (producing sediment that accumulates in the subduction zone), newer magmas become increasingly silicic and become intermediate. During oceanic-continental collisions (Fig. 18), the generally mafic magma rises through felsic continental lithosphere to build a volcanic arc on the continental margin. Assimilation-melting of the overlying felsic continental plate produces intermediate magma. The different magma types and their relation to plate movements are illustrated in figure (19).

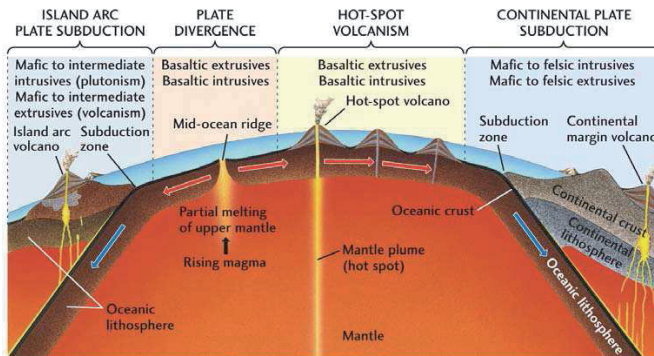


Fig. 19. Magma types vs. plate movements.

- Types of Granites**

The granites could be classified based on mineralogy, geochemistry and tectonic emplacement:

- ✓ Mineralogical classifications (IUGS classification) (Fig. 20).
- ✓ Chemical classification (alumina saturation, S-I-A-M classification etc.) (Fig. 21).
- ✓ Tectonic classification (based on plate tectonic setting) (Fig. 22).

SIAM classification:

S-type Granitoid

Chapter 1: Basic Geologic Concepts

- derived due to partial melting of sedimentary and metasedimentary rock.
- more common in collision zones.
- peraluminous granites [i.e., $\text{Al}_2\text{O}_3 > (\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO})$] and have $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio < 0.3 .
- characterised by muscovite, biotite and marginally higher SiO_2 contents

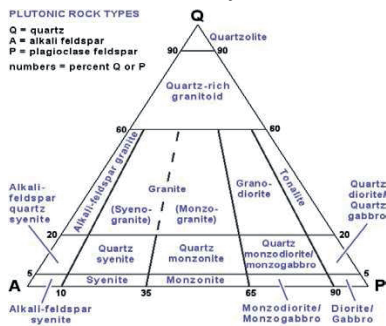


Fig. 20. IUGS classification of Granites.

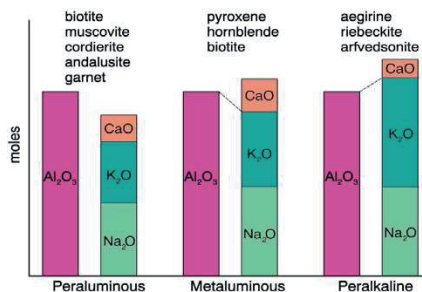


Fig. 21. Alumina saturation classes based on the molar proportions of $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$ ("A/CNK").

I-type Granitoid

- derived due to partial melting of igneous proloith.
- derived from igneous or metaigneous rocks of lower continental crust subjected to partial melting due to upwelling of mantle material to higher levels.
- generally metaluminous granites, expressed mineralogically by the absence of peraluminous minerals like muscovite (with exceptions) and have $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio > 0.3 .
- characterised by presence of hornblende/alkali amphiboles $\pm$ biotite.

M-type Granitoid (Fig. 22)

- Derived due to fractional crystallisation of basaltic magma.

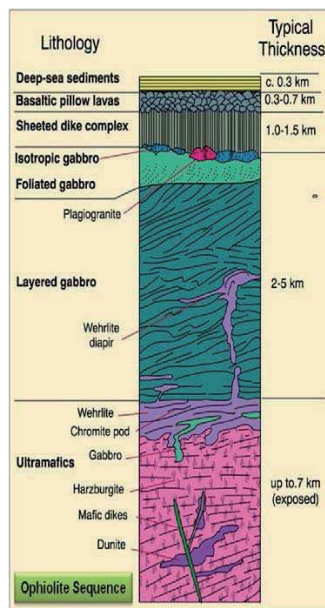


Fig. 22. Ophiolite sequence.

Chapter 1: Basic Geologic Concepts

- Relatively Plagioclase rich (plagiogranite of ophiolite).
- Associated with Gabbros and Tonalites in the field.
- Formed in subduction zone.

A-type Granitoid (anorogenic type)

- emplaced in either within plate anorogenic settings or in the final stages of an orogenic event.
- High SiO₂ (~73.81%)
- High F contents (6000 to 8000 ppm)
- Presence of fluorite is an important characteristic of A-type granites.

Table 2: Characteristics of SIAM Granitoids.

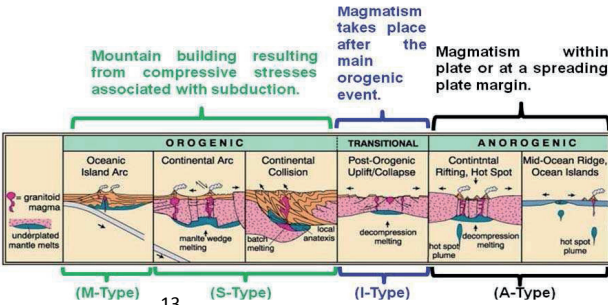
SIAM Characteristics										
Table 18-3. The S-I-A-M Classification of Granitoids										
Type	SiO ₂	K ₂ O/Na ₂ O	Ca, Sr	Al/(C+N+K)*	Fe ³⁺ /Fe ²⁺	Cr, Ni	δ ¹⁸ O	⁸⁷ Sr/ ⁸⁶ Sr	Misc	Petrogenesis
M	46-70%	low	high	low	low	low	< 9‰	< 0.705	Low Rb, Th, U Low LIL and HFS	Subduction zone or ocean-intraplate
I	53-76%	low	high in mafic rocks	low; metaluminous to peraluminous	moderate	low	< 9‰	< 0.705	high LIL/HFS med Rb, Th, U hornblende magnetite	Mantle-derived Subduction zone Intracrustal
S	65-74%	high	low	high metaluminous	low	high	> 9‰	> 0.707	variable LIL/HFS high Rb, Th, U biotite, cordierite Als, Grt, Ilmenite	Mafic to intermed. igneous source Subduction zone
A	high → 77%	Na ₂ O high	low	var peralkaline	var	low	var	var	low LIL/HFS high Fe/Mg high Ga/Al High REE, Zr High F, Cl	Supracrustal sedimentary source Anorogenic Stable craton Rift zone

* molar Al₂O₃/(CaO+Na₂O+K₂O)

Data from White and Chappell (1983), Clarke (1992), Whalen (1985)

Based on Tectonic emplacement, granitoids occur in areas where the continental crust has been thickened by orogeny, either continental arc subduction or collision (Fig. 23). The majority of granitoids are derived by crustal anatexis, however, mantle may also be involved. The mantle contribution may range from that of a source of heat for crustal anatexis, or it may be the source of material as well.

Fig. 23. Tectonic emplacement of different SIAM granitoids.



• Ophiolite sequence

Ophiolites consist of five distinct layers.

- The first layer is the youngest and is primarily sediment that was accumulated on the seafloor.
- The second layer is pillow basalt. Pillow basalt is characterized by large pillow. When erupting lava encounters the cold sea water, the outside of the lava immediately crystallizes, forming a thick crust. The extremely hot lava still inside the blob, oozes out of the crust and instantly crystallizes again.
- The next layer consists of sheeted dikes. Sheeted dikes form by rising magma within the earth's crust. As the sheeted dikes cool fractures and cracks occur in the rock.
- Gabbro underlays sheeted dikes and compositionally similar to basalt. Isotropic (massive) gabbro, indicates fractionation of magma chamber. Layered gabbro, resulting from settling out of minerals from a magma chamber.
- The bottommost layer is peridotite, which is believed to be mantle rock composition.

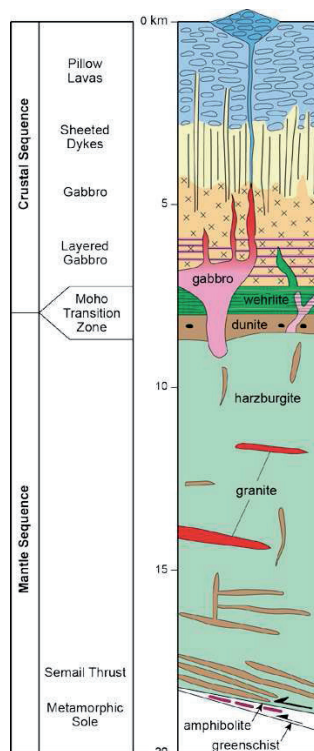


Fig. 24. Ophiolite sequence.

4- Metamorphism

It is a process leading to changes in mineralogy and/or texture in a rock (Fig. 25). The boundary between diagenesis and metamorphism defines by noting the first occurrence of a mineral that does not occur as a detrital or diagenetic mineral in surface sediments, (e.g. chlorite, epidote, lawsonite, laumontite, albite, zeolite, etc). Formation of some of these minerals requires a temperature of at least 150-

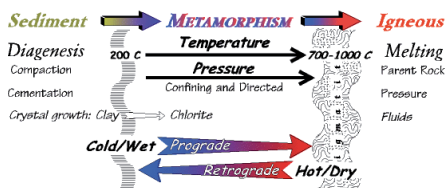


Fig. 25. Metamorphism and diagenesis.

200 °C or 1500 bars or depth of about 5 km under normal geothermal conditions. The upper limit of metamorphism is defined as the beginning of appreciable melting.

- **Agents of metamorphism**

1. Heat is the most important source of energy allowing the formation of new and more stable mineral and textural reconstruction and recrystallization during metamorphism.
2. Pressure (measured in bars - 1 kb is approximately each 3 km depth) changes both a rock's mineralogy and its texture. Pressure comes in different varieties; confining pressure, directed pressure (or stress), burial pressure and fluid pressure.
3. Chemically Active Fluids (ion transport): In some metamorphic settings, new materials are introduced by the action of hydrothermal solutions (hot water with dissolved ions). Many metallic ore deposits form in this way.

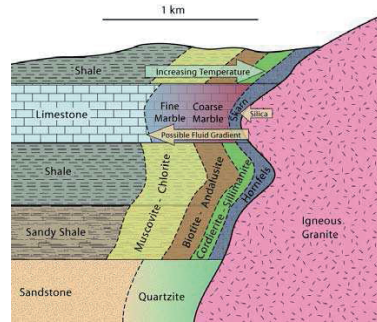


Fig. 26. Contact metamorphism.

- **Types of metamorphism**

1. Contact metamorphism occurs when magma invades cooler rock. Here, a zone of alteration called an aureole (or halo) forms around the emplaced magma (Fig. 26). These large aureoles often consist of distinct zones of metamorphism. Near the magma body, high temperature minerals such as garnet may form, whereas farther away such low-grade minerals as chlorite are produced. Contact metamorphism produces a zone of alteration called an aureole around an intrusive igneous body. Shales baked by igneous contact form very hard fine-grained rocks called Hornfels. Calcareous rocks (dirty limestones) when

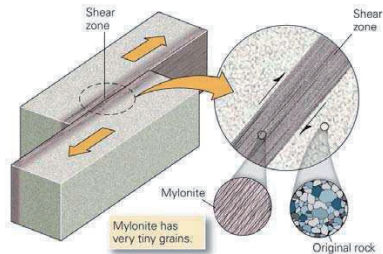


Fig. 27. Dynamic metamorphism.

subject to contact metamorphism an alteration by hot fluids produce rocks called Skarns. Pyrometamorphism: Very high temperatures at very low pressures, generated by a volcanic or subvolcanic body.

2. Metamorphism along fault zones is known as dynamic metamorphism (Fig. 27). In some cases, rock may even be milled into very fine components. The result is a loosely coherent rock called fault breccia that is composed of broken and crushed rock fragments. This type of localized metamorphism, which involves purely mechanical forces that pulverize individual mineral grains, is called cataclastic metamorphism. Much of the intense deformation associated with fault zones occurs at great depth. In this environment the rocks deform by ductile flow, which generates elongated grains that often give the rock a foliated or lineated appearance. Rocks formed in this manner are termed mylonites.

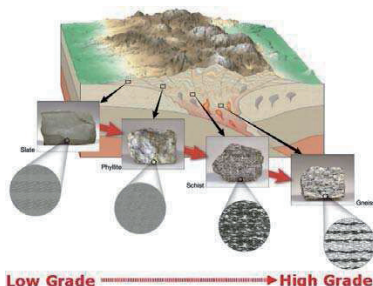


Fig. 28. Regional metamorphism.

3. Regional Metamorphism. The metamorphic rocks produced during regional metamorphism are associated with mountain building (orogenic metamorphism – convergent plate boundaries). During these dynamic events, large segments of Earth's crust are intensely squeezed and become highly deformed (Fig. 28). As the rocks are folded and faulted, the crust is shortened and thickened, like a rumpled carpet. This general thickening of the crust results in terrains that are lifted high above sea level. In regional metamorphism, there usually exists a gradation in intensity. As we shift from areas of low-grade metamorphism to areas of high grade metamorphism, changes in mineralogy and rock texture can be observed.

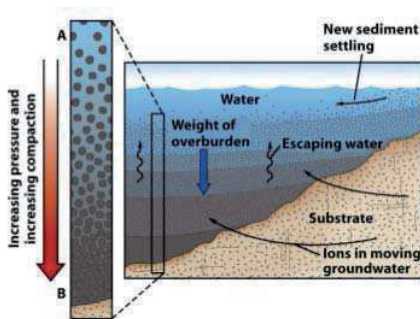


Fig. 29. Burial metamorphism.

4. Burial metamorphism. Metamorphic effects are attributed to increased pressure and temperature due to burial (Fig. 29). Range from diagenesis to the formation of zeolites, prehnite, pumpellyite, laumontite, etc. Diagenesis and lithification start when rocks reach several kilometers depth. Continued burial leads to low grade burial metamorphism. It is common for sedimentary structures in the unaltered rocks to remain in the metamorphosed rocks, indicating relatively little recrystallization. This style of metamorphism grades into regional metamorphism with increasing pressure and temperature. We find it in deep sedimentary basins.

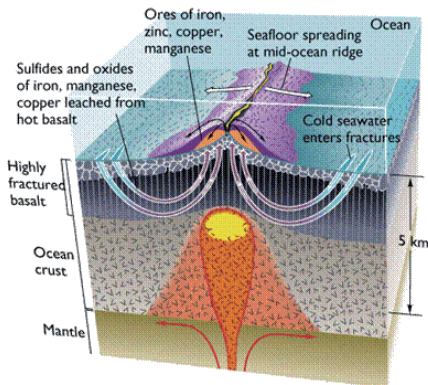


Fig. 30. Hydrothermal metamorphism.

5. High-pressure low- temperature metamorphism: This metamorphism is associated with subduction zones. It is called high pressure/low temperature metamorphism where the subducting plates has been cooled by interaction with seawater.
6. Hydrothermal metamorphism: It is caused by hot H₂O-rich fluids and usually involving metasomatism (Fig. 30). This style of metamorphism is distinguished by high fluid content and changes in rock composition. It occurs when hot water percolates (or convects) through rock. This happens around plutons and in association with

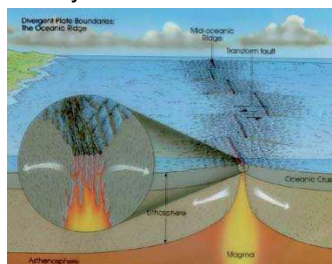


Fig. 31. Ocean-floor metamorphism.

underwater volcanism. Pressures are usually low and temperatures moderate. By dissolving components that are least compatible within the rocks, hydrothermal metamorphism can produce very exotic deposits. Sulfides and massive ore bodies are associated with it.

7. Ocean-Floor Metamorphism: It affects the oceanic crust at ocean ridge spreading centers (Fig. 31). May be considered another example of hydrothermal metamorphism. Highly altered chlorite-quartz rocks-distinctive high-Mg, low-Ca composition. Metamorphic rocks exhibit considerable metasomatic alteration, notably loss of Ca and Si and gain of Mg and Na. These changes can be correlated with exchange between basalt and hot seawater.

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1- What is "Economic Geology"?

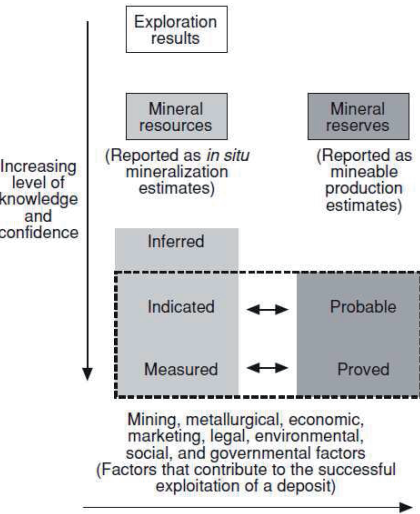
The discipline of “Economic Geology” covers all aspects pertaining to the description and understanding of mineral resources. Ore deposits are formed when a useful commodity is sufficiently concentrated in an accessible part of the Earth’s crust so that it can be profitably extracted. Ore deposits are natural concentrations of useful metals, minerals or rocks, which can be economically exploited. Concentrations that are too small/low-grade for mining are called occurrences or mineralizations.

Fe, Al, Mg, Ti, and Mn, are abundantly distributed in the Earth’s crust (i.e. between about 0.5 and 10 wt%) and only require a relatively small degree of enrichment in order to make a viable deposit. Table 1 shows that Fe and Al, for example, need to be concentrated by factors of 9 and 4 respectively, relative to average crustal abundances, in order to form potentially viable deposits. The crustal abundances for Au and Pt are in the range 4–5 parts per billion (ppb) and even though ore deposits routinely extract these metals at grades of around 5 g t⁻¹, the enrichment factors involved are between 1000 and 1250 times.

Table 1 Average crustal abundances for selected metals and typical concentration factors that need to be achieved in order to produce a viable ore deposit

	Average crustal abundance	Typical exploitable grade	Approximate concentration factor
Al	8.2%	30%	×4
Fe	5.6%	50%	×9
Cu	55 ppm	1%	×180
Ni	75 ppm	1%	×130
Zn	70 ppm	5%	×700
Sn	2 ppm	0.5%	×2500
Au	4 ppb	5 g t ⁻¹	×1250
Pt	5 ppb	5 g t ⁻¹	×1000

Note: 1 ppm is the same as 1 g t⁻¹.



Mineral deposits are basically valuable rocks. Their formation is compared with processes that have produced ordinary rocks and is

Fig. 1. Simplified scheme illustrating the conceptual difference between mineral resources and ore reserves as applied to mineral occurrences.

investigated with petrological methods. Mineral deposits can also be thought of as a geochemical enrichment of elements or compounds in the Earth's crust, which is determined by their chemical properties. Exploration results can be translated into a mineral resource once it is clear that an occurrence of intrinsic economic interest exists in such form and quantity that there are reasonable prospects for its eventual exploitation (Fig. 1). Such a resource can only be referred to as an ore reserve if it is a part of an economically extractable measured or indicated mineral resource.

The purpose of this process-orientated course is to provide a better understanding of the nature and origin of mineral occurrences and how they fit into the Earth system.

2- Classification scheme of "Ore Deposits"

A very simple classification of ores is achieved on the basis of igneous, sedimentary/surficial and hydrothermal categories. All mineral deposits can be classified into three types based on process, namely magmatic deposits, hydrothermal deposits and surficial deposits formed by surface and groundwaters. Ore-forming processes can overlap between igneous and hydrothermal and between sedimentary and hydrothermal (Fig. 2).

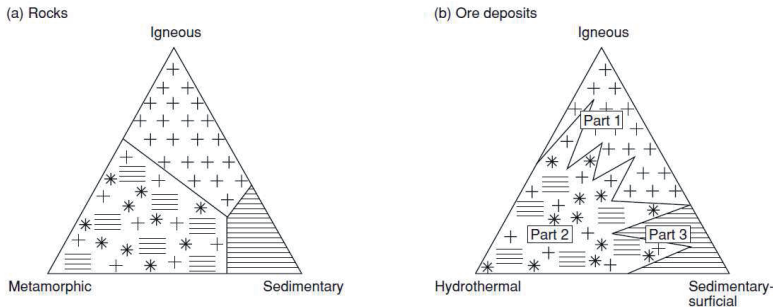


Fig. 2. Classification of the principal rock types (a) and an analogous, but much simplified, classification of ore deposit types (b).

Various geological aspects are employed to classify ore deposits (Table 2), including:

1. The presence of certain metals or minerals (e.g. silver, haematite);
2. The form of the orebody (vein, bed, etc.);

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3. The local geological environment (submarine or terrestrial volcanism);
4. The plate tectonic setting (island arc, continental margin) and
5. Other genetic characteristics such as formation temperatures and fluid chemistry.

However, a stringent genetic classification of mineral deposits is very difficult. One reason for this is that many ore deposits represent a position in a complex multi-dimensional space of well defined end members:

1. The formation of Kuroko ore deposits, for example, is an interplay of volcanic, intrusive, sedimentary and diagenetic processes;
2. The origin of high-grade BIF-haematite ore seems to comprise sedimentation induced by proliferating marine life, later passage of saline basinal brines and supergene components.

Table 2. Genetic classification of Ore Deposits.

<i>I. Magmatogenic Ore Deposits</i>	
1. Orthomagmatic Deposits:	
Sulphide Fe-Ni-(Cu-PGE) ore hosted by Archaean komatiites and subvolcanic ultramafic intrusions; Alpine type Cr-PGM in ophiolites, and seams in layered mafic intrusions; Cu-Ni-PGM "reefs" in layered mafic intrusions; complex mafic-ultramafic intrusions with, for example, conduit-hosted Cu-Ni-PGM; impact magma bodies with Ni-Cu-PGM; Ural-Alaska type ultramafic ring intrusions with Cr-PGM; Ti-Fe in Mesoproterozoic anorthosite-ferrodiorite complexes; orthomagmatic iron ore deposits of iron oxides and apatite in intermediate to felsic igneous rocks (Kiruna type); apatite-Fe-Nb-Zr, or REE-U-F in carbonatite plugs and nephelinite intrusions	
2. Pegmatites with ore of Be, Li, Rb, Cs, Ta (Nb), U, Th, REE, Mo, Bi, Sn and W, industrial minerals, gemstones	
3. Magmatic-Hydrothermal Deposits:	
Skarn ore, with magnetite-Cu-Co-Au, W, Zn-Pb-Ag, Mo-Bi-Au and Sn-As-Pb-Zn-W-Mo; contact-metasomatic ore (Pb, Ag, Zn); Fe-oxide-Cu-Au (U-REE) deposits (IOCG); porphyry deposits (Cu-Mo-Au, Sn-W); submarine volcanogenic (Kuroko) and volcanic-hosted massive sulphide deposits (VMS); vein deposits (Sn, W, Cu); terrestrial epithermal Ag-base metal deposits; epithermal Au-Ag deposits	
<i>II. Supergene Ore Deposits</i>	
1. Residual (Eluvial) Deposits:	
Residual placers (e.g. W, Sn); bauxite; lateritic Au, Fe and Mn ore deposits	
2. Supergene Enrichment Deposits:	
Enriched sulphide ore (Cu, Ag); lateritic Ni	
3. Infiltration Deposits	
U in sandstone; Pb-Zn-F-Ba and Mn in karst cave systems	
<i>III. Sedimentary Ore Deposits</i>	
1. Allochthonous:	
Colluvial, alluvial (gold, columbite, cassiterite, wolframite, platinum) and coastal (rutile, ilmenite, zircon, monazite) placers	
2. Autochthonous:	
Sulphide deposits, mainly in black shales; polymetallic deposits of Cu-Sb-Zn-Pb-Ag (-Au), mostly of sedex type; Palaeoproterozoic banded Fe ore (BIF) of Algoma and Superior type, and banded Mn ore; oolitic Fe and Mn ore; deep sea manganese nodules and crusts (Mn-Cu-Ni-Co-PGM)	
<i>IV. Diagenetic-Hydrothermal Ore Deposits</i>	
1. Stratabound and/or stratiform sediment-hosted Cu deposits:	
European Copper Shale (Cu); Central African Copper Belt (Cu, Co, Pb, Zn, U)	
2. Mississippi Valley type (MVT) Pb-Zn-F-Ba deposits (hosted in marine carbonates)	
3. Saline brine-related deposits	
Pb-Zn-F-Ba, metasomatic siderite, pre-concentration of high-grade hematite	
<i>V. Metamorphosed and Metamorphic Ore Deposits</i>	
Metamorphism of pre-existing ore generally improves processing characteristics of ore, but is rarely a factor of metal accumulation and ore formation; metamorphic examples include $\pm$ <i>in situ</i> redistribution, concentration and recrystallization of gold	
<i>VI. Metamorphogenic-Hydrothermal Ore Deposits</i>	
Prograde and retrograde metamorphogenic-hydrothermal ore deposits (e.g. orogenic Au deposits in accretion-subduction-collision complexes; part of the Central African Cu-Co ore deposits; Cu in Mt Isa, Australia).	

3- Important definitions

Ore: A type of rock that contains minerals with important elements including metals that can be extracted from the rock at a profit (Fig. 3).



Fig. 3. Iron Ore.

Ore deposits: Parts of the crust, where ores are concentrated (Fig. 4).

Gangue: Commercially worthless material that surrounds, or is closely mixed with, a wanted mineral in an ore deposit (Fig. 5).

Mining: Extraction of ores, or other valuable minerals from the ore deposits (Fig. 6).

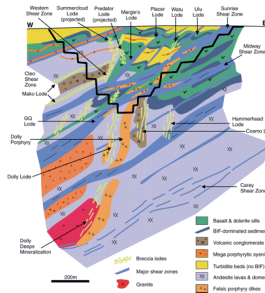


Fig. 4. Ore deposit.

Metallogeny: The study of the genesis of mineral deposits, with emphasis on their relationships in space and time to geological features of the Earth's crust.

Metallotect: any geological, tectonic, lithological or geochemical feature that has played a role in the concentration of one or more elements in the Earth's crust.



Fig. 5. Gangue.

Metallogenic Epoch: a unit of geologic time favorable for the deposition of ores or characterized by a particular assemblage of deposit types.

Metallogenic Province: a region characterized by a particular assemblage of mineral deposit types.



Fig. 6. Gangue.

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Syngenetic: refers to ore deposits that form at the same time as their host rocks (Fig. 7).

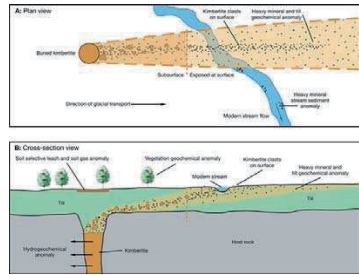


Fig. 7. Syngenetic ore formation.

Epigenetic: refers to ore deposits that form after their host rocks.

Hypogene: refers to mineralization caused by ascending hydrothermal solutions.

Supergene: refers to mineralization caused by descending solutions (Fig. 8).

Endogenetic: concentration caused by processes in the Earth's interior (magmatism or metamorphism) (Fig. 9).

Exogenetic: concentration caused by processes at the Earth's surface (sedimentation, weathering).

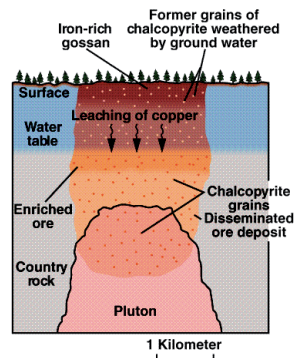


Fig. 8. Supergene enrichment.

Lateral secretion: concentration of metals by abstraction from surrounding rock.

Hypothermal: hydrothermal ore deposits formed at substantial depths (greater than 4500 meters) and elevated temperatures (400–600 °C).

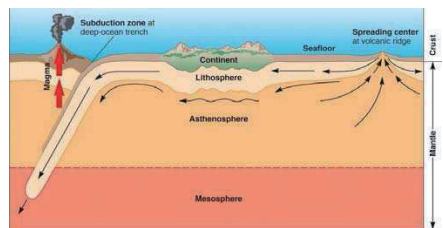


Fig. 9. Endogenetic processes.

Mesothermal: hydrothermal ore deposits formed at intermediate depths (1500–4500 meters) and temperatures (200–400 °C).

Epithermal: hydrothermal ore deposits formed at shallow depths (less than 1500 meters) and fairly low temperatures (50–200 °C).

4- Common ore and gangue minerals

It is estimated that there are about 3800 known minerals that have been identified and classified. Only a very small proportion of these make up the bulk of the rocks of the Earth's crust, as the common rock forming minerals. Likewise, a relatively small number of minerals make up most of the economically viable ore deposits of the world. The following compilation is a breakdown of the more common ore minerals in terms of chemical classes based essentially on the anionic part of the mineral formula. The compilation also includes some of the more common "gangue," which are those minerals that form part of the ore body, but do not contribute to the economically extractable part of the deposit.

The compilation, including ideal chemical formulae, is subdivided into six sections, these are: native elements, halides, sulfides and sulfo-salts, oxides and hydroxides, oxy-salts (such as carbonates, phosphates, tungstates, sulfates) and silicates. These groups are described as following:

- **Native elements**

Both metals and non-metals exist in nature in the native form, where essentially only one element exists in the structure. Copper, silver, gold, and platinum are all characterized by cubic close packing of atoms, have high densities, and are malleable and soft. The carbon atoms in diamond are linked in tetrahedral groups forming well cleaved, very hard, translucent crystals. Sulfur occurs as rings of eight atoms and forms bipyramids or is amorphous. Examples of native metals are: Gold – Au; Silver – Ag; Platinum – Pt; Palladium – Pd and Copper – Cu. Examples of native non-metals are: Sulfur – S; Diamond – C and Graphite – C.

- **Halides**

The halide mineral group comprises compounds made up by ionic bonding. Minerals such as halite and sylvite are cubic, have simple chemical formulae, and are highly soluble in water. Halides sometimes form as ore minerals, such as Halite – NaCl; Sylvite – KCl; Chlorargyrite – AgCl; Fluorite – CaF₂ and Atacamite – Cu₂Cl(OH)₃.

- **Sulfides and sulfo-salts**

This is a large and complex group of minerals in which bonding is both ionic and covalent in character. The sulfide group has the general formula AMXP, where X, the larger atom, is typically S but can be As, Sb, Te, Bi, or Se, and A is one or more metals. The sulfo-salts, which are much rarer than sulfides, have the general formula AMBNXP, where A is commonly Ag, Cu, or Pb, B is commonly As, Sb, or Bi, and X is S. The sulfide and sulfo-salt minerals are generally opaque, heavy and have a metallic to sub-metallic luster.

Sulfides include: Chalcocite – Cu₂S; Bornite – Cu₅FeS₄; Galena – PbS; Sphalerite – ZnS; Chalcopyrite – CuFeS₂; Pyrrhotite – Fe_{1-x}S; Pentlandite – (Fe,Ni)₉S₈; Millerite – NiS; Covellite – CuS; Cinnabar – HgS; Skutterudite – (Co,Ni)As₃; Sperrylite – PtAs₂; Cobaltite – CoAsS; Gersdorffite – NiAsS; Loellingite – FeAs₂; Molybdenite – MoS₂; Realgar – AsS; Orpiment – As₂S₃; Stibnite – Sb₂S₃; Bismuthinite – Bi₂S₃; Argentite – Ag₂S; Calaverite – AuTe₂; Pyrite – FeS₂; Laurite – RuS₂; Braggite/cooperite – (Pt,Pd,Ni)S and Moncheite – (Pt,Pd)(Te,Bi)₂

Sulfo-salts include: Tetrahedrite – (Cu,Ag)₁₂Sb₄S₁₃; Tennantite – (Cu,Ag)₁₂As₄S₁₃ and Enargite – Cu₃AsS₄

- **Oxides and hydroxides**

This group of minerals is variable in its properties, but is characterized by one or more metal in combination with oxygen or a hydroxyl group. The oxides and hydroxides typically exhibit ionic bonding. The oxide minerals can be hard, dense, and refractory in nature (magnetite, cassiterite) but can also be softer and less dense, forming as products of hydrothermal alteration and weathering (hematite, anatase, pyrolucite). Hydroxides, such as goethite and gibbsite, are typically the products of extreme weathering and alteration.

Oxides include: Cuprite – Cu₂O; Hematite – Fe₂O₃; Ilmenite – FeTiO₃; Hercynite – FeAl₂O₄; Gahnite – ZnAl₂O₄; Magnetite – Fe₃O₄; Chromite – FeCr₂O₄; Rutile – TiO₂; Anatase – TiO₂; Pyrolucite – MnO₂; Cassiterite – SnO₂; Uraninite – UO₂; Thorianite – ThO₂ and Columbite-tantalite – (Fe,Mn)(Nb,Ta)₂O₆

Hydroxides (or oxyhydroxides) include: Goethite – $\text{FeO}(\text{OH})$; Gibbsite – $\text{Al}(\text{OH})_3$; Boehmite – $\text{AlO}(\text{OH})$ and Manganite – $\text{MnO}(\text{OH})$

- **Oxy-salts**

The carbonate group of minerals form when anionic carbonate groups $(\text{CO}_3)^{2-}$ are linked by intermediate cations such as Ca, Mg, and Fe. Hydroxyl bearing and hydrated carbonates can also form, usually as a result of weathering and alteration. The other oxy-salts, such as the tungstates, sulfates, phosphates, and vanadates, are analogous to the carbonates, but are built around an anionic group of the form $(\text{XO}_4)^n$. Carbonates include: Calcite – CaCO_3 ; Dolomite – $\text{CaMg}(\text{CO}_3)_2$; Ankerite – $\text{CaFe}(\text{CO}_3)_2$; Siderite – FeCO_3 ; Rhodochrosite – MnCO_3 ; Smithsonite – ZnCO_3 ; Cerussite – PbCO_3 ; Azurite – $\text{Cu}_3(\text{OH})_2(\text{CO}_3)_2$ and Malachite – $\text{Cu}_2(\text{OH})_2\text{CO}_3$.

Tungstates include: Scheelite – CaWO_4 and Wolframite – $(\text{Fe},\text{Mn})\text{WO}_4$.

Sulfates include: Baryte(s) – BaSO_4 ; Anhydrite – CaSO_4 ; Alunite – $\text{KAl}_3(\text{OH})_6(\text{SO}_4)_2$; Gypsum – $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$ and Epsomite – $\text{MgSO}_4 \cdot 7\text{H}_2\text{O}$

Phosphates include: Xenotime – YPO_4 ; Monazite – $(\text{Ce},\text{La},\text{Th})\text{PO}_4$ and Apatite – $\text{Ca}_5(\text{PO}_4)_3(\text{F},\text{Cl},\text{OH})$.

Vanadates include: Carnotite – $\text{K}_2(\text{UO}_2)(\text{VO}_4)_2 \cdot 3\text{H}_2\text{O}$.

- **Silicates**

The bulk of the Earth's crust and mantle is made up of silicate minerals that can be subdivided into several mineral series based on the structure and coordination of the tetrahedral $(\text{SiO}_4)^{4-}$ anionic group. Silicate minerals are generally hard, refractory and translucent. Most of them cannot be regarded as ore minerals in that they do not represent the extractable part of an ore body, and the list provided below shows only some of the silicates more commonly associated with mineral occurrences as gangue or alteration products. Some silicate minerals, such as zircon and spodumene, are ore minerals and represent important sources of metals such as zirconium and lithium, respectively. Others, such as kaolinite, are mined for their intrinsic properties (i.e. as a clay for the ceramics industry).

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Nesosilicates include: Zircon – $\text{Zr}(\text{SiO}_4)$; Garnet (almandine) – $\text{Fe}_3\text{Al}_2(\text{SiO}_4)_3$; Garnet (grossular) – $\text{Ca}_3\text{Al}_2(\text{SiO}_4)_3$; Sillimanite – Al_2SiO_5 ; Topaz – $\text{Al}_2\text{SiO}_4(\text{F},\text{OH})_2$ and Chloritoid – $(\text{Fe},\text{Mg},\text{Mn})_2(\text{Al},\text{Fe})\text{Al}_3\text{O}_2(\text{SiO}_4)_2(\text{OH})_4$.

Sorosilicates include: Lawsonite – $\text{CaAl}_2\text{Si}_2\text{O}_7(\text{OH})_2\cdot\text{H}_2\text{O}$ and Epidote – $\text{Ca}_2(\text{Al},\text{Fe})_3\text{Si}_3\text{O}_{12}(\text{OH})$.

Cyclosilicates include: Beryl – $\text{Be}_3\text{Al}_2\text{Si}_6\text{O}_{18}$ and Tourmaline – $(\text{Na},\text{Ca})(\text{Mg},\text{Fe},\text{Mn},\text{Al})_3(\text{Al},\text{Mg},\text{Fe})_6\text{Si}_6\text{O}_{18}(\text{BO}_3)_3(\text{OH},\text{F})_4$.

Inosilicates (chain) include: Tremolite-actinolite – $\text{Ca}_2(\text{Fe},\text{Mg})_5\text{Si}_8\text{O}_{22}(\text{OH})_2$; Spodumene – $\text{LiAlSi}_2\text{O}_6$ and Wollastonite – CaSiO_3 .

Phyllosilicates (sheet) include: Kaolinite – $\text{Al}_4\text{Si}_4\text{O}_{10}(\text{OH})_8$; Montmorillonite – $(\text{Na},\text{Ca})_{0.3}(\text{Al},\text{Mg})_2\text{Si}_4\text{O}_{10}(\text{OH})_2\cdot n\text{H}_2\text{O}$; Illite – $\text{KAl}_2(\text{Si},\text{Al})_4\text{O}_{10}(\text{H}_2\text{O})(\text{OH})_2$; Pyrophyllite – $\text{Al}_2\text{Si}_4\text{O}_{10}(\text{OH})_2$; Talc – $\text{Mg}_3\text{Si}_4\text{O}_{10}(\text{OH})_2$; Muscovite – $\text{KAl}_2(\text{AlSi}_3\text{O}_{10})(\text{OH})_2$; Biotite – $\text{K}(\text{Fe},\text{Mg})_3(\text{Al},\text{Fe})\text{Si}_3\text{O}_{10}(\text{OH},\text{F})_2$; Lepidolite – $\text{K}(\text{Li},\text{Al})_3(\text{Si},\text{Al})_4\text{O}_{10}(\text{OH},\text{F})_2$ and Chlorite – $(\text{Fe},\text{Mg},\text{Al})_5\text{--}6(\text{Si},\text{Al})_4\text{O}_{10}(\text{OH})_8$.

Tectosilicates include: Quartz – SiO_2 ; Orthoclase – $(\text{K},\text{Na})\text{AlSi}_3\text{O}_8$; Albite – $(\text{Na},\text{Ca})\text{AlSi}_3\text{O}_8$; Scapolite – $(\text{Na},\text{Ca})_4[(\text{Al},\text{Si})_4\text{O}_8]_3(\text{Cl}, \text{CO}_3)$ and Zeolite (analcime) – $\text{NaAlSi}_2\text{O}_6\cdot\text{H}_2\text{O}$.

Unknown structure includes: Chrysocolla – $(\text{Cu},\text{Al})_2\text{H}_2\text{Si}_2\text{O}_5(\text{OH})_4\cdot n\text{H}_2\text{O}$

Some applications of ore minerals are shown in Table 3.

Table 3. Applications of some ore minerals.

Metallic Elements	Ore Minerals	Chemical Formulae	Industrial Usage
Copper (Cu)	Chalcocopyrite	CuFeS ₂	Alloys, Electronics
	Bornite	Cu ₅ FeS ₄	
Gold (Au)	Native gold	Au	Precious metals, electronics, chemical instruments
	Electrum	AuAg	
Lead (Pb)	Galena	PbS	Batteries, alloys, glasses
Nickel (Ni)	Pentlandite	(Fe, Ni) ₉ S ₈	Special steel alloys, rockets, nuclear reactors
Molybdenum (Mo)	Molybdenite	MoS ₂	Special steel, filaments, glass pigments
Platinum (Pt)	Native platinum	Pt	Catalysts, Electronics, Chemical instruments
Mercury (Hg)	Cinnabar	HgS	Electric industrie, Catalysts, corrosives
Zinc (Zn)	Sphalerite	ZnS	Alloys, pesticides, medicines
Antimony (Sb)	Stibnite	Sb ₂ S ₃	tin tubings, bronze, enamel, ceramics
	Tetrahedrite	Cu ₁₂ Sb ₄ S ₁₃	
Aluminum (Al)	Gibbsite	Al(OH) ₃	Alloys, automobiles, aircrafts
Uranium (U)	Uraninite	UO ₂	Nuclear fuels, catalysts, pigments
Silver (Ag)	Native silver	Ag	Precious metals, alloys, photos, electric plating
	Argentite	Ag ₂ S	
Tin (Sn)	Cassiterite	SnO ₂	Tin plates, bronze
tungsten (W)	Sheelite	CaWO ₄	Special steels, ultralight macbinaries, alloys
	Wolframite	(Fe, Mn)WO ₄	
Cobalt (Co)	Linnaeite	Co ₃ S ₄	Steel alloys, ceramics, catalysts
Chrome (Cr)	Chromite	(Fe, Mg)Cr ₂ O ₄	Alloys, plating, refractory bricks, dyes
Titanium (Ti)	Rutile	TiO ₂	High pressure vessels, textiles, dyes

5- Metals and minerals for a high-tech world

The availability of metal, non-metal and mineral raw materials, particularly those that underpin high-technology sectors, is important for the ongoing development of many industries (Figs. 10 and 11). Major ore – commodities - such as iron ore, coal, aluminium and copper are very



Fig. 10. Minerals for cellular phones.

important in a wide range of sectors, however there is a diversity of supply and substantial resources. In essence a mineral resource, i.e., ore deposit is critical if it is both economically important and has high risk of supply disruption. These supply risks originate from four main causes: (1) scarcity of the ore or

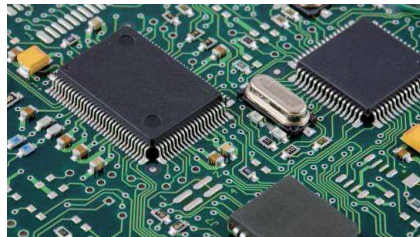


Fig. 11. Minerals for circuits.

mineral (the geological abundance); (2) diversity and stability of supply; (3) production only as a by-product of other commodities; and (4) level of

concentration of ore production and processing within particular countries or by particular companies.

6- Uses of critical and other metal and mineral commodities

The periodic table of the elements illustrates the groupings of elements with certain shared physical and chemical properties. For example, all metals are good conductors of electricity and are generally malleable and ductile, whereas semi-metals are semi-conductors of electricity, a highly valuable property in electronics (Fig. 12) and solar energy panels (Fig. 13). Some sub-groups have particular shared properties, for example platinum-group elements (including platinum and palladium) and other noble metals such as gold are highly resistant to chemical corrosion.



Fig. 12. Minerals for electronics.



Fig. 13. Minerals for electronics.

Other metals are valued for their extremely high melting temperatures and hardness, such as tungsten and rhenium, so that alloys of these metals tend to have greater tensile strength at high temperatures. This property enables rhenium-bearing super-alloys in jet engine turbine blades to operate at higher temperatures than non-rhenium turbines, reducing aeroplane emissions and fuel costs (Fig. 14). The rare-earth elements, which include the lanthanide series metals as well as scandium and yttrium, have diverse and very useful properties. For example, small percentages of neodymium and dysprosium in some alloys increase permanent magnet strength by orders of



Fig. 14. The use of rhenium in high temperature turbines in the aerospace industry.

magnitude, enabling step changes in miniaturizing of telecommunications and other electronic devices, and much more efficient generation of electricity in commercial wind turbines.

7- Mineral Deposit versus Orebody

A mineral deposit/ore deposit may be defined as a rock body that contains one or more elements (or minerals) sufficiently above the average crustal abundance to have potential economic value. Mineral deposits can be classified into two broad categories:

1. metallic mineral deposits (e.g., deposits of copper, lead, zinc, iron, gold, etc.), from which one or more metals can be extracted; and
2. nonmetallic (or industrial) mineral deposits (e.g., deposits of clay, mica, fluorite, asbestos, garnet, etc.), which contain minerals useful on account of their specific physical or chemical properties.

8- Styles of Mineralization and Morphology of Mineral Deposits

The style of mineralization refers to the pattern of distribution of ore minerals in a host rock, and it varies from being very subtle (even invisible to the naked eye as in some precious metal deposits) to quite pronounced (as in the case of massive sulfide deposits). The shapes of mineral deposits are also highly variable, from concordant tabular and stratiform to discordant veins and breccia bodies.

Table 4. Morphology of Ore Deposits.

	Mode of occurrence	Typical examples
<i>Disseminated</i>	Ore minerals dispersed through the host rock(Fig. 15)	Diamond in kimberlite pipes
<i>Stockwork</i>	An interlacing network of small and narrow (commonly measured in centimeters), close-spaced ore-bearing veinlets traversing the host rock (Fig. 16)	Footwall alteration zone of volcanic-hosted massive sulfide deposits (see Figs. 10.1, 10.4)
<i>Massive</i>	Mineralization comprising >50% of the host rock(Fig. 17)	Volcanic-hosted massive sulfide lenses (see Fig. 10.4)
<i>Tabular</i>	An ore zone that is extensive in two dimensions, but has a restricted development in its third dimension(Fig. 18)	Sandstone-type uranium deposits (see Figs. 14.3, 14.7)
<i>Vein-type</i>	Mineralization in veins, commonly discordant to the host rock layering (depositional)(Fig. 19)	Base- and precious metal veins (see Figs. 2.28, 2.29)
<i>Stratiform</i>	Mineralization confined to a specific bed and, thus, broadly conformable to the host rock layering (depositional)(Fig. 20)	Kupferschiefer-type stratiform copper deposits (see Fig. 12.4)
<i>Strata-bound</i>	Mineralization discordant to host rock layering (depositional), but restricted to a particular stratigraphic interval	Mineralized breccia bodies in Mississippi Valley-type deposits (see Fig. 13.9)

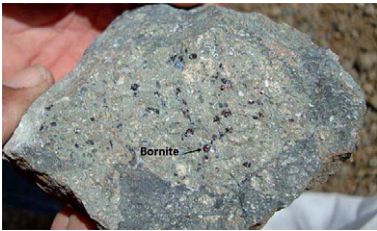


Fig. 15. Disseminated Ore Deposit.



Fig. 16. Stockwork Ore Deposit.

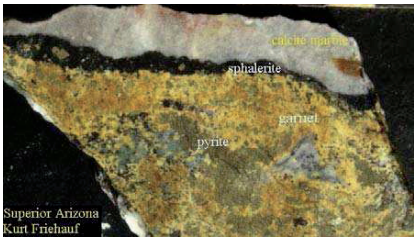


Fig. 17. Massive Ore Deposit.



Fig. 18. Tabular Ore Deposit.



Fig. 19. Vein Ore Deposit.

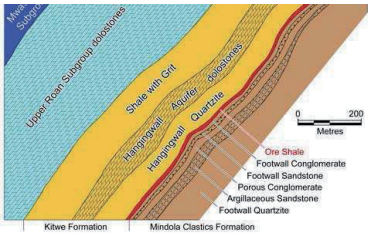


Fig. 20. Stratiform Ore Deposit.

Chapter 3: Magmatic Ore Deposits

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1- Ore forming processes.

All the common ore-forming elements are present in magmas and ordinary rocks, in amounts ranging from a few parts per billion to several thousands of parts per million. Selective concentration of one or more ore constituents to form a mineral deposit is achieved by some combination of the following:

1. extraction of the constituents from magmas, rocks, and oceans;
2. transport of the constituents in a fluid medium from the source region to the site of deposition; and
3. localization of the constituents at certain favorable sites.

The ore-forming processes may be grouped into the following four broad categories:

1. Orthomagmatic processes
2. Sedimentary processes
3. Metamorphic processes
4. Hydrothermal processes

2- Orthomagmatic Processes.

Orthomagmatic ore-forming processes are related to the evolution of magmas emplaced at crustal levels (Fig. 1). The two end members of this span continuum processes are:

1. orthomagmatic processes – resulting in concentration of ore minerals as a direct consequence of silicate melt magmatic crystallization; and
2. (magmatic) hydrothermal processes – leading to concentration of ore minerals from magmatic hydrothermal fluids by crystallization (because of the small quantity of dissolved water, crystallization of mafic and ultramafic magmas seldomly leads to the generation of large amounts of ore-forming hydrothermal fluids, except perhaps when substantial assimilation of water-bearing crustal rocks is involved, i.e, rare hydrothermal fluids result after the crystallization of mafic magmas).

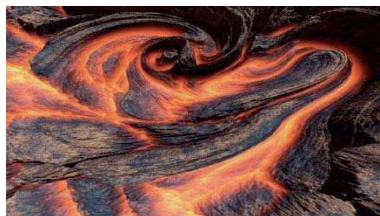


Fig. 1. Magma.

Deposits of iron, copper, nickel, chromium, titanium, and platinum, are restricted to mafic and ultramafic rocks. In addition, deposits of some of these

metals characteristically occur in particular kinds of mafic and ultramafic rocks

- e.g.,

1. chromium in dunite and peridotite,
2. nickel in peridotite and norite, and
3. titanium in gabbro and anorthosite.

A genetic relationship between felsic magmas and mineral deposits is much less convincing, because the association of metals with specific felsic rocks is not as clear as with mafic and ultramafic rocks. Of the deposits commonly associated with felsic intrusives, only those of tin are restricted to granites (Fig. 2). Other deposits – such as those of copper,



Fig. 2. Tin Ore.

silver, gold, lead, zinc, molybdenum, tungsten – are associated with rocks ranging from granite to diorite, although there may be a preferential association with a particular rock type in a given geologic setting. On the other hand, the well-established tendency of mineral deposits to cluster near the periphery of felsic intrusives and metal zoning centered on such intrusives strongly suggest a genetic connection between felsic magmas and the associated hydrothermal deposits (Fig. 3).

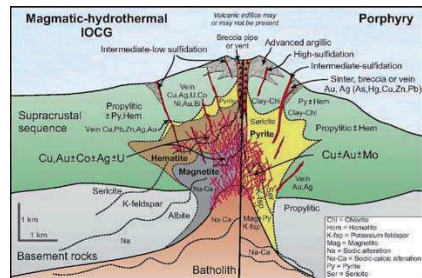


Fig. 3. Hydrothermal deposits the periphery of felsic intrusive.

• Magmas as sources of ore constituents

Magmas are generated by partial melting of lower crustal or upper mantle material. Magmas are essentially silicate melts with variable amounts of ore metals and other elements such as water and relatively minor amounts of other volatile constituents (e.g., CO₂, H₂S, SO₂, HCl, HF, H₂). Partial melting of the top 100-200 km of the upper mantle by adiabatic decompression (pressure-release melting) produces primary magmas of mafic (basaltic or picritic) or ultramafic (komatiitic) composition in most tectonic settings (Fig. 4).

The two main end-member models of partial melting are (Fig. 5):

- a) equilibrium or batch melting that involves continuous reaction and equilibration of the partial melt with the crystalline residue (solid mantle in our case), until mechanical conditions allow the melt to escape (or segregate) as a single “batch” of magma; and
- b) fractional melting in which the partial melt is continuously removed from the system as soon as it is formed, thereby preventing further reaction between the melt and the solid residue (solid mantle in our case).

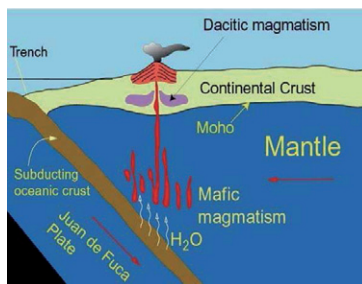


Fig. 4. Adiabatic decompression of upper mantle and upwelling of magma.

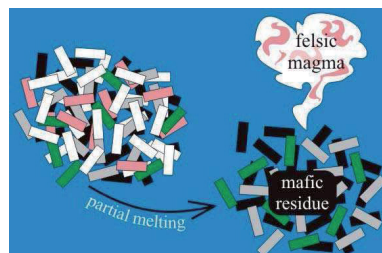


Fig. 5. Partial melting.

• Sulfur in Magmas

Sulfur is one of the most abundant volatiles in magmas. Sulfur has significant effects on the partitioning of a wide variety of elements between silicate melts, liquid metals, gases, and solids, and consequently magmatic sulfur species exert major controls on the genesis of a large variety of ore deposits. The behavior of sulfur in silicate melts/hydrothermal solutions is much more complex than that of other volatiles, such as water and carbon dioxide, because of its different oxidation states. At low oxygen fugacities (concentration), sulfide (S^{2-}) is the predominant sulfur species whereas at higher oxygen fugacities sulfate (SO_4^{2-}) is dominant. Other species such as sulfite (S_4^{+}) may exist as well at specific conditions. It is often difficult to predict the behavior of sulfur.

The mantle, with an estimated sulfur concentration in the range of 300-1,000 ppm, is believed to be the dominant source of sulfur carried in basaltic magmas. During partial melting of the mantle the available iron sulfide would melt well before the beginning of silicate melting.

The sulfur concentration in oceanic basalts is from 600 ± 150 ppm to as high as 1,600 ppm. It is, however, difficult to predict the sulfur contents of silicate melts, because the solubility of sulfur is controlled by a number of interdependent variables, such as temperature, pressure, O₂, S₂ and, especially, the activities of FeO and SiO₂ in the melt. The sulfur solubility in silicate melts decreases with:

1. decreasing temperature,
2. increasing activity of FeO or increasing activity of SiO₂, and
3. decreasing S₂ or increasing O₂.

The actual amount of juvenile sulfur (liquid sulfur) carried by a basaltic magma might be significantly higher than its saturation limit at the source, if some of the sulfide melt in a given volume of mantle material was incorporated into the partial melt as an immiscible phase. The sulfur content might also be enhanced by assimilation of sulfur from the country rocks.

I-type granitoid magmas have a greater potential for bulk assimilation of country-rock sulfur than S-type magmas (Fig. 6).

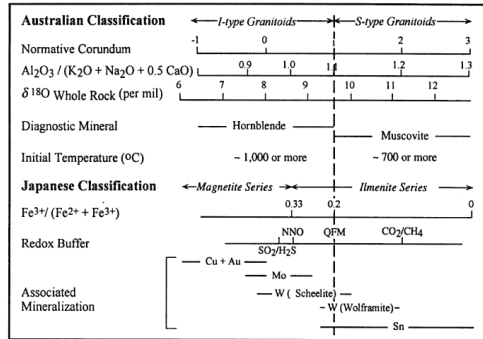


Fig. 6. I-type vs. S-type granites.

• Water in Magmas

The generation of significant amounts of water-saturated magmas or hydrous fluids is unlikely in the upper mantle because of its low water content. On the other hand, dioritic and granitic magmas generated by partial melting of lower crustal rocks are likely to be more hydrous and capable of generating an aqueous fluid phase with progressive crystallization (magmatic hydrothermal solutions).

The separation of liquid phase/hydrothermal fluid (aqueous/vapor) from a magma, is controlled mainly by the solubility of H₂O in the melt, which is very strongly pressure dependent but, however, only weakly temperature dependent.

Chapter 3: Magmatic Ore Deposits

The amount of hydrothermal fluid that will be exsolved from magma depends on its initial H₂O content, its depth of emplacement, and its crystallization history (Fig. 7). The initial H₂O contents of magmas ranges from “2.5 to 6.5 wt%”, with a median value close to 3.0 wt% (in basaltic magma). For dioritic and granitic magmas, the initial melt would contain in excess of 3.3 wt% H₂O. When an ascending water-bearing magma begins to crystallize, the volume of the residual magma becomes smaller and smaller, and H₂O (with other volatiles) gets concentrated in this decreasing volume. The exsolved aqueous hydrothermal fluid phase can be highly saline.

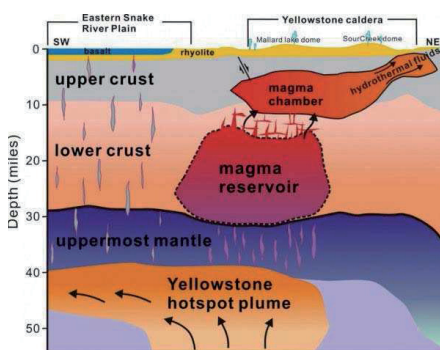


Fig. 7. Exsolved hydrothermal solution.

The sulfur content of the aqueous fluid/hydrothermal solution is determined by its SO₂:H₂S ratio that increases with increasing O₂ of the parent magma. Aqueous fluids/hydrothermal solutions derived from I-type magmas (with high O₂) may contain large quantities of SO₂ as well as H₂S. However, at lower temperatures/cooling the hydrolysis of SO₂ ($4\text{SO}_2 + 4\text{H}_2\text{O} = \text{H}_2\text{S} + 3\text{H}_2\text{SO}_4$) and/or the reaction with Fe²⁺-bearing minerals of the wallrock ($\text{SO}_2 + 6\text{FeO} + \text{H}_2\text{O} = \text{H}_2\text{S} + 3\text{Fe}_2\text{O}_3$); the activity of H₂S increases, causing precipitation of sulfide ore minerals from the metal chloride complexes in the hydrothermal solution.

• Concentration of ore minerals by magmatic crystallization

Ore constituents present in magma may be concentrated further during the course of crystallization. Three magmatic differentiation processes have been considered particularly important for the formation of orthomagmatic ore deposits: liquid immiscibility; gravitative crystal settling and filter pressing.

- i. **Liquid Immiscibility:** Liquid immiscibility is the phenomenon of separation of a cooling magma into two or more liquid phases of different composition in equilibrium with each other.

There are three cases of liquid immiscibility under geologically reasonable conditions (Fig. 8):

1. separation of Fe-rich tholeiitic magmas into two liquids, one felsic (rich in SiO_2) and the other mafic (rich in Fe);
2. splitting of CO_2 -rich alkali magmas into one melt rich in CO_2 and the other rich in alkalis and silica, which may account for the origin of carbonatite magmas (alkaline); and
3. segregation of sulfide melts (or oxysulfide melts containing a few percent dissolved oxygen) from sulfide-saturated mafic or ultramafic magmas.

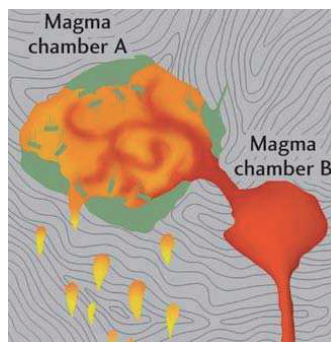


Fig. 8. Exsolved hydrothermal solution.

Conditions or processes that are likely to promote sulfide immiscibility in a mafic or ultramafic magma are:

- a) cooling of the magma, which not only decreases its sulfur solubility, but also causes crystallization of silicate minerals, thereby increasing the sulfur concentration in the residual magma;
 - b) silica enrichment of the magma by reaction with felsic country rocks
 - c) mixing of a more fractionated magma with a less fractionated magma, both of which were nearly saturated with sulfur; and
 - d) assimilation of sulfur from country rocks.
- e) Other processes which can, in theory, cause sulfide saturation are oxidation and an increase in pressure.

Fractional segregation typically occurs during the crystallization of a sulfide-saturated silicate magma, because the crystallization of even a small amount of olivine (or other sulfur-free minerals) leads to sulfide immiscibility. A small amount of sulfide melt segregating from a silicate magma is likely to be dispersed as minute droplets (more dense) in the magma. Chalcophile elements (e.g., Ni, Cu) are strongly partitioned into the sulfide melt (Fig. 9).

Sulfide immiscibility induced by a sudden change in intensive parameters (e.g., due to sulfur or silica assimilation from country rocks) should produce batch segregation of sulfide melt. Such sulfide segregation may or not be

accompanied by silicate crystallization, but sulfide segregation before the onset of significant silicate crystallization would provide a more favorable situation for the formation of magmatic segregation deposits.

- ii. Gravitational Settling: The formation of massive deposits of magmatic crystallization products, such as chromite and sulfides, requires that they are concentrated by some mechanism in a restricted part of the magma chamber. A possible mechanism of crystal-liquid separation in a magma undergoing crystallization is gravitational settling (or floating) of crystals by virtue of their density differences relative to the liquid (Figs. 10, 11). Cumulate layers, including chromite rich layers, in large differentiated complexes such as the Bushveld and the Stillwater, have generally been regarded as products of gravitational crystal settling.

- iii. Filter Pressing: Magmatic segregation deposits may also form by crystallization of residual magmas. A mafic magma without a high enough O₂ for early crystallization of Fe-Ti oxide minerals would produce enrichment of iron and titanium in the residual magma. This heavier liquid, then, may drain downward, collect below as a

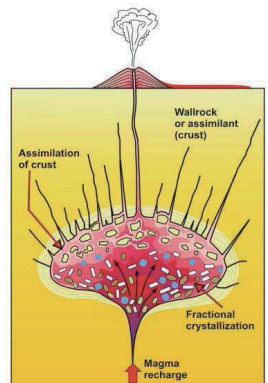
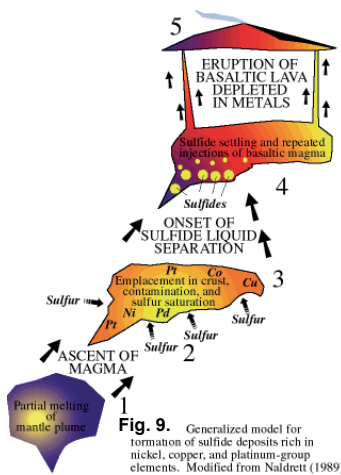


Fig. 10. Gravitational settling.

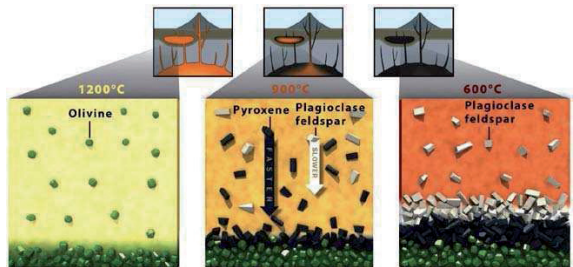
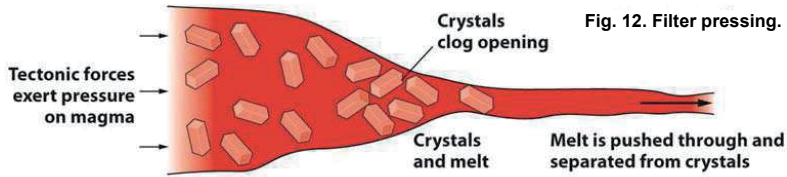


Fig. 11. Gravitational settling.

segregation resting on a solid floor of early formed sunken crystals, and crystallize into a layer with significant concentration of Fe-Ti oxide minerals. In some situations, the residual magma may be squeezed out by filter pressing and form magmatic injection deposits (Fig. 12). The Fe-Ti oxide deposits associated with anorthosites and anorthositic gabbros are believed to have formed by gravitative accumulation and injection of residual magmas.



The Magmatic Ore Formation Systems involve:

1. Orthomagmatic ore formation
2. Ore deposits at mid-ocean ridges and in ophiolites
3. Ore formation related to alkaline magmatic rocks, carbonatites and kimberlites
4. Granitoids and ore formation processes
5. Ore deposits in pegmatites
6. Hydrothermal ore formation
7. Skarn- and contact-metasomatic ore deposits
8. Porphyry copper (Mo-Au-Sn-W) deposits
9. Hydrothermal-metasomatic ore deposits
10. Hydrothermal vein deposits
11. Volcanogenic ore deposits

2.1 Orthomagmatic ore formation

2.1.1 Mafic-Ultramafic Complexes: Chromium, Nickel Copper and Platinum group elements (PGE)

Oxides (magnetite, ilmenite, chromite), base metal sulphides (Ni, Cu), and ore of precious metals (Pt, Pd, Au) is often found in ultramafic and mafic igneous rocks. These ores were formed at magmatic temperatures, while the melt was essentially liquid and before total solidification of the magma. Therefore, this class of ore deposits is called “orthomagmatic”. Enrichment processes concentrate/segregate low metal traces from a large mass of silicate melt into small volumes. However, a common evolution is that the parent melt evolves towards saturation so that either a solid (e.g. chromite) or a liquid (e.g. sulphide melt) accumulates the metal.

Because of their higher density compared to the inheriting silicate liquids, ore melt droplets or solid ore phases typically cumulates below still liquid magma (gravitational accumulation/segregation) (Fig. 13). Consolidation of cumulate minerals can lead to expulsion of inter-cumulus liquid (filter pressing). As the system (magma) cools, ore melts may separate into cumulates (e.g. Fe-sulphides) and residual liquids (Cu-rich sulphide melt).

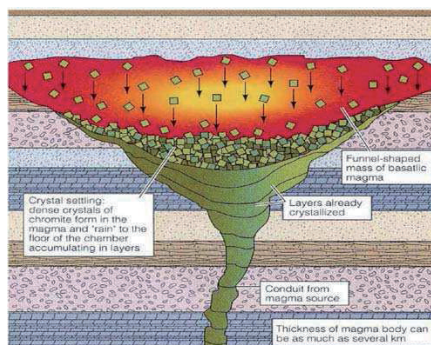


Fig. 13. Gravitational accumulation/segregation of chromite

Concentration of metals such as PGM (platinum group metals), Au, Ni and Cu in sulphide melt is controlled by the Nernst partition coefficient (D) between sulphide and silicate liquids, and by other kinetic factors. In addition, a disequilibrium is controlled by silicate/sulphide liquid mass ratio “R-factor”.

A zone refining model is appropriate when for example, sulphide droplets sink through a magma chamber and collect chalcophile metals (Ag, As, Bi, Cd, Cu, Ga, Ge, Hg, In, Pb, Po, S, Sb, Se, Sn, Te, Tl and Zn). This is followed by resorption of iron-sulphide liquid in under-saturated

magma leading to concentration of limited base metal (Ni, Cu, Zn,...) together with very high content of PGM (Pt, Pd) and precious metals (Au) enrichment.

Most orthomagmatic ore deposits are found in intrusive rocks. Gravitational settling can explain many features of ore formation in layered mafic intrusions (Fig. 14). Often, the formation and segregation of a sulphide melt, enriched with metal, - outside/far from the silicate melt - is the key to enrichment of exploitable metals. Volcanic/eruptive equivalents are also notable, such as the Ni-Cu-Fe sulphides in komatiitic lava flows (Fig. 15), or the magnetite and haematite lavas and tuffs in andesitic-rhyolitic volcanoes. (komatiite is a type of ultramafic mantle-derived volcanic rock with high to extremely high Mg content).



Fig. 14. Ores in layered ultramafic/mafic intrusions.



Fig. 15. Komatiite.

The orthomagmatic ore bodies are layers in stratified magmatic rocks (often formed as cumulates), lenses or cross-cutting dykes and veins. This depends on the morphology of the segregation (sedimentation) surface and on dynamic factors during ore formation. Massive ore is the product of highly efficient unmixing of ore particles or melt droplets and silicates, whereas disseminated mineralization reflects lower efficiency. Highly complex ore body shapes can be found in flow channels and pipes of mafic lavas.

Examples of orthomagmatic ores are:

- a) Cr-PGE deposits at Bushveld Igneous Complex, South Africa,
- b) Ni-Cu-PGE deposits at The Great Dykes, Zimbabwe,
- c) Ni-PGE-Cr deposits at Sudbury "(meteorite impact-unusual), Canada,
- d) Ni-Cu-PGE deposits at Stillwater Igneous Complex, Montana, US.

• **Bushveld complex**

It is largest preserved layered intrusion in the world. Bushveld Complex, in South Africa, is hosting an exceptional variety and mass of high grade metal ores. The Bushveld Intrusive Complex comprises the layered mafic-ultramafic intrusion which contains enormous metal resources. These mafic layers are overlapped by granites containing host less important fluorite and tin ores.

Bushveld complex consists of ultramafic-mafic sequence which reaches a thickness of 9000 m. It is strongly layered. The major units from bottom to top comprise (Fig. 16):

1. the Lower Zone with dunite, bronzitite, and harzburgite;
2. the conspicuously banded Critical Zone (Fig. 17) with a lower part of orthopyroxenite, chromitite bands and some harzburgite, and a higher part marked by the first cumulus plagioclase and by cyclic layering of economically significant platiniferous

chromitite, harzburgite, bronzitite, norite and anorthosite in this order (cyclic units); its upper boundary is marked by the Merensky Reef (Pt, Ni, Cu);

3. the Main Zone with gabbronorite and minor layering;
4. the Upper Zone with magnetite (ferro) gabbro and ferrodiorite, which contains numerous magnetite (V-Ti) layers.

There is no consensus of opinion on the number, nature, volume and source of the different magma types and the plate setting for magmatism of Bushveld complex. One opinion is the occurrence of cratonic extensional associated with strike-slip movement (Fig. 17). The occurrence of A-type granites,

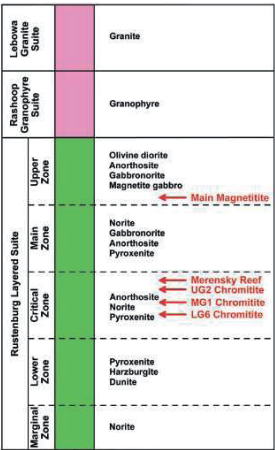


Fig. 16. Generalized stratigraphic column of the Bushveld Complex. Red arrows refer to the position of ore bearing layers.



Fig. 16. Generalized stratigraphic column of the Bushveld Complex. Red arrows refer to the position of ore bearing layers.

which are generally associated with crustal extension, is consistent with this hypothesis.

The volume of magma formed the Bushveld suggests the interaction of a mantle plume with lithosphere that has been thinned to between 110 and 50 km (Fig. 17). A hot Lower Zone magma derived from a mantle diapir which halted in the lower crust, flattening of the diapir led to the melting of the lower crust and the formation of the lower Critical Zone magma. During the accumulation of the Lower and Critical Zones, the magma chamber was continually fed by olivine- and orthopyroxene- crystallizing magmas that formed the Lower and Critical Zones (Fig. 18).

Progressive mixing of new and residual fractionated magma (Fig. 18) resulted in the slow evolution from a harzburgite/orthopyroxenite dominated Lower Zone, through a feldspathic orthopyroxenite dominated lower Critical Zone, to a norite/anorthosite dominated upper Critical Zone.

In general, layered mafic intrusions occur in several geodynamic settings:

1. Archaean greenstone belts;
2. intracratonic regions (the Bushveld Complex);
3. at passive margins of continents; and (Fig. 19)

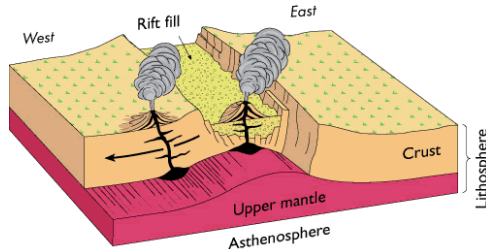


Fig. 17. cratonic extensional associated with strike-slip movement.

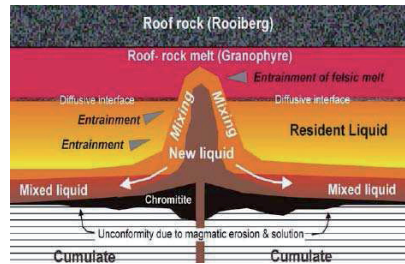


Fig. 18. Continuous magmatic feed to form the Bushveld Complex.

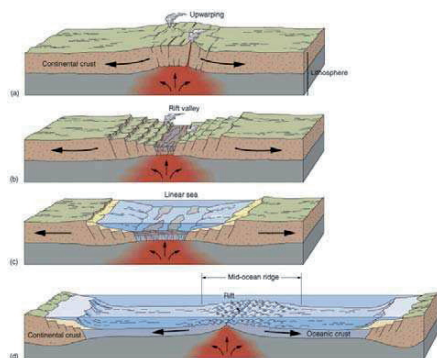


Fig. 19. Opening a rift valley.

4. in active orogenic belts.

Intracratonic regions that experienced tensional tectonics can also exhibit unstratified, very complex mafic-ultramafic intrusions with Cu-Ni PGM ores.

- **Sudbury: Impact magma bodies with orthomagmatic ore deposits**

Mineralized impact structures are very rare. A giant example is the Sudbury Igneous Complex (SIC) of Ontario, Canada, the second largest source of nickel+copper+platinum in the world (Fig. 20).

The SIC is the remnant of a voluminous melt body that has been produced by the impact of a meteorite into continental crust (Fig. 21). Ore deposits occur mainly in embayments of the footwall contact of the intrusion, in radiating dykes “offsets” and within intensely brecciated footwall rocks up to 2km from the contact.

Total past production and current reserves of the Sudbury District are estimated at >1700Mt of Ni, Cu, Co, Pt, Pd, Au and Ag ore. Among approximately 90 known Ni-Cu-PGE deposits, 14 are currently worked.

At Sudbury, lithologic zonation is interpreted to be due to gravity separation of mafic and felsic liquids that formed an emulsion immediately after the impact (Fig. 20). The ore-bearing sublayer displays typical features of mafic cumulates and gravity segregation of sulphide liquids. Offset dykes and footwall deposits host an important part of metal resources.

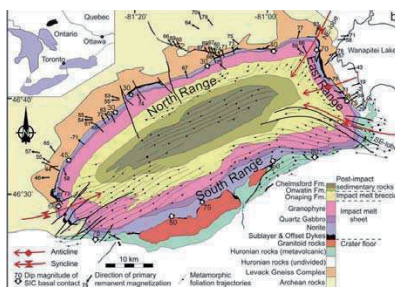


Fig. 20. Overview map of the Sudbury impact structure, Canada, one of the giant nickel-copper mining districts of the world.

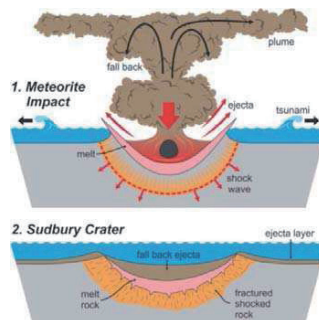


Fig. 21. Sudbury meteoric impact.

2.1.2 Anorthosite-ferrodiorite complexes

Many rocks contain small amounts of titanium locked in silicate minerals (e.g., biotite, amphibole), but the economically found in anorthosites as Ti-rich oxide minerals (Fe-Ti oxides, magnetite and ilmenite-hematite solid solution series) and Ti-oxides (mainly rutile). Anorthosite is an intrusive igneous rock characterized by a predominance of plagioclase feldspar (90–100%), and a minimal mafic component (0–10%) (Fig. 22).

Orebodies consist of ilmenite and/or rutile, magnetite or haematite, and a gangue of apatite and some graphite. The anorthosites are commonly coarsely crystalline, rather massive than layered and consist of >90wt.% andesine to labradorite. Anorthosite plutons may be associated with coeval intrusions of, ferrogabbro and ferrodiorite.

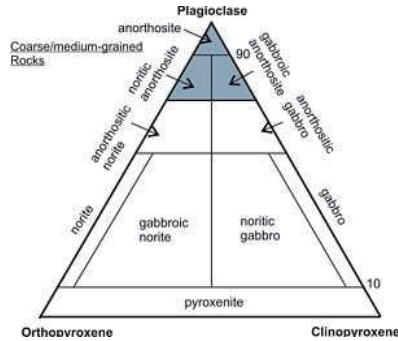


Fig. 22. Anorthosite.

Resulting ore bodies are stratiform and either massive or disseminated (Sanford Lake (New York, USA) and Lac Tio (Quebec, Canada). From anorthosite rocks (Fig. 23), 50% of the world's titanium supply is derived; they also contain about half of the total titanium resources.



Fig. 23. Rutile in anorthosite.

It is believed that anorthosites are the products of basaltic magma after the “mechanical removal” of mafic minerals. Since the mafic minerals are not found with the anorthosites, these minerals must have been left at either a deeper level or the base of the crust. A typical theory is as follows: partial melting of the mantle generates a basaltic magma, which does not immediately ascend into the crust. Instead, the basaltic magma forms a

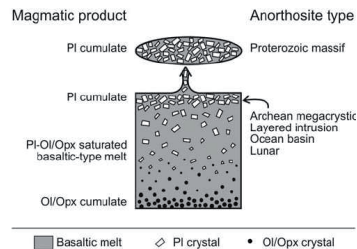


Fig. 24. Schematic model elaborating the formation of anorthosite.

large magma chamber at the base of the crust and fractionates large amounts of mafic minerals, which sink to the bottom of the chamber (Fig. 24). The co-crystallizing plagioclase crystals float, and eventually are emplaced into the crust as anorthosite plutons (Fig. 25). Most of the sinking mafic minerals form ultramafic cumulates which stay at the base of the crust.

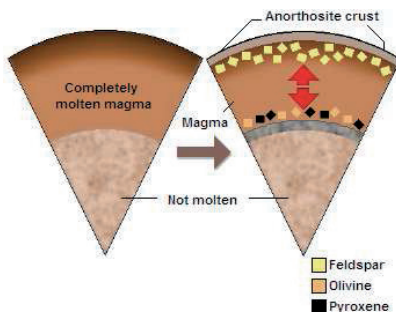


Fig. 25. Emplacement of the anorthosite plutons.

2.1.3 Fe-rich melts segregated from intermediate to felsic magmas

The metallogeny of Fe-ore segregated from intermediate to acidic melt is ambiguous case of orthomagmatic ore formation. Although it is possible that FeOx rich melt would separate from acidic magma when the acidic magma is enriched in O₂ there is no general agreement that this is a path to the formation of large ore deposits.

This debate is attributed to the difficulty of segregating Fe-ores by gravity in high viscosity of SiO₂-rich magma (intermediate/granitic). However, such segregation is possible when:

- The magma is sheared by slow convection so that the low-viscosity FeOx liquid may be concentrated; and
- the possible high content of sodium and phosphorous acts as fluxing agents for iron melt.

Mineral segregation under these conditions would produce ore of magnetite and apatite in the proportion of about 2 : 1, as exploited in the Kiruna District (Sweden). High fluorine and chlorine content of the apatites, and the presence of minerals such as amphibole and scapolite, imply an important role of magmatic volatiles (H₂O, Cl, F, CO₂, etc.) which promote segregation and mobility of ore melt (Fig. 26).

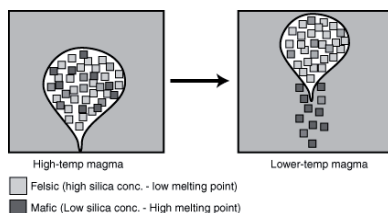


Fig. 26. segregation and mobility of ore melt.

Kiruna in northern Sweden, is considered as the largest iron ore of orthomagmatic origin in felsic intrusions, because the ore is co-genetic with the host rocks trachyandesite and rhyodacite. Lower Ti and V concentrations distinguish this type of iron ore - in felsic intrusions - from massive iron oxides segregated from mafic magmatic melts. As well, this type of Fe-ores (in Kiruna) is also characterized by lack Cu and Au when compared with that formed by hydrothermal solutions.

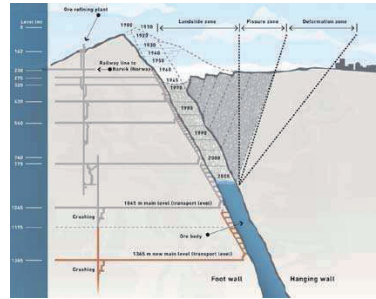


Fig. 27. Fe-ore in Kiruna, northern Sweden.

An extrusive origin is also considered for magnetite ore bodies at El Laco, Chile (Fig. 28). Magnetite or haematite-apatite ores have been described as massive and vesicular lavas, veins, crystal tuffs and pyroclastic agglomerates deposited by volcanoes built of rhyolite.



Fig. 28. Fe-ore in an extrusive at El Laco, Chile.

In conclusion, orthomagmatic deposits of iron oxides and apatite in intermediate to felsic igneous rocks (intrusive and extrusive types) may originate by mixing and mingling of ultra-mafic and silicic melt.

2.2 Ore deposits at mid-ocean ridges and in ophiolites

Exploration of ocean floors resulted not only in the recognition of plate tectonics (Fig. 29) but also in the discovery of conspicuous signs of active ore forming systems – the “black and white smokers” and ophiolites.

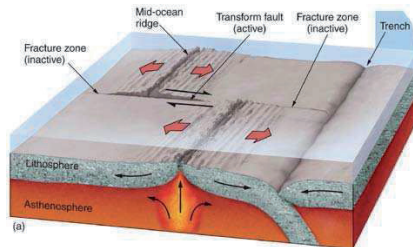


Fig. 29. Plate tectonics and hot spots.

2.2.1 Ores in Ophiolites

Ophiolites are fragments of oceanic crust and mantle (Fig. 30) that have been transported (obducted) as thrust sheets (nappes) towards continental masses. The tectonic emplacement was normally associated with dismemberment of the original succession.

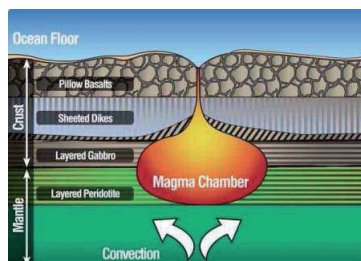


Fig. 30. Ophiolite sequence.

A complete ophiolite sequence comprises (Fig. 30):

1. Extrusive basalts of typical chemical (MORB) characteristics at the top, often in the shape of pillow lavas; ocean floor metamorphism of basalt increases from the zeolite facies at the top to greenschist facies at the bottom;
2. The sheeted dyke complex, consisting of vertical basalt dykes, many ophiolites, however, lack sheeted dykes;
3. The plutonic complex, comprising higher intrusive homogeneous gabbro, diorite, tonalite and trondhjemite ("plagiogranite"), and deeper layered gabbro and peridotites, that display properties of cumulate rocks (the "cumulate sequence"); the magmatic rocks are normally not metamorphosed;
4. The tectonized and depleted mantle, dominated by large masses of serpentinite (after harzburgite) and characteristic pods of dunite.

Tectonized (foliated) harzburgite and the lower cumulates host dunite bodies that may contain massive and disseminated chromite ore. Dunite in harzburgite can be understood as lag segregation from rising basaltic melt diapirs. Chromitites originate from dunite by liquid-liquid immiscibility. Because of ductile shearing in the oceanic mantle, both dunites and chromite orebodies are strongly deformed, resulting in lenticular pod-like shapes.

2.2.2 Black smokers

Submarine black smoker vents are hydrothermal cones or chimneys that may reach a height of about 20 m, built on outcrops of basalt (Fig. 31). Black smokers are sea vents "geysers" that occur on the ocean floor and spew hot, mineral-rich water, that help support a diverse community of organisms (Fig.

32). From an opening at the top, a high speed jet of hot fluid is ejected. The vents are tubes with zoned walls, from pyrite and chalcopyrite inside through sphalerite, marcasite, barite, anhydrite and amorphous SiO_2 to the exterior (Fig. 33).

Oxidation of sulphides by seawater “seafloor weathering” produces vari-coloured ochreous alteration fragments, which mainly consist of iron oxy-hydroxides that assemble on the sea floor around the vents and build gossan-like mounds (Gossan is oxidised surficial sulphide deposits). The expulsion temperature of the metalliferous solutions is 350°C . The hot Na-Ca-Cl fluids of the black smokers are reducing and have pH from 4–5, salinities from 0.1 to 3 times seawater, elevated iron, copper, zinc, barium and SiO_2 , and traces of As, Cd, Li, Be, Cs, Mn, B, Cl, HCl, H_2S , and CH_4 .

Different solutes are derived from various protoliths (Fig. 34), possibly from magma, and reflect also different conditions of water/rock reactions. For example, copper is enriched relative to iron under moderately oxidizing conditions, whereas a low O_2 results in a high Fe/Cu ratio. If iron prevails, black or grey smoke-like plumes of amorphous iron sulphide and iron-manganese oxy hydroxides rise several hundred metres upwards and disperse over a distance of many kms. When zinc is concentrated in the fluids the smokers are bluish.

Fluid properties change by phase separation, boiling, alteration and mineral precipitation during rise to the seafloor. Upon discharge at the ocean floor, hot acidic fluids

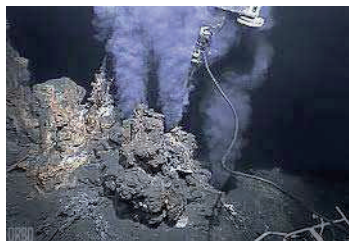


Fig. 31. Black smokers.



Fig. 32. Black smokers and mineral-rich water.



Fig. 33. Cross section in a Black smokers vent.

mix with cold alkalic seawater, which results in immediate precipitation of solutes (Fig. 34).

2.2.3 White smokers

White smoker vents discharge fluids between 100 and 300°C (Fig. 35). They form mainly:

- a) in the early stage of a newly established hydrothermal system; and
- b) by sub-seafloor mixing of hot black smoker fluid with cooler waters.

The second probably leads to precipitation of sulphides at depth. Therefore, white smokers may indicate the presence of hidden stockwork and vein deposits of copper and zinc. SiO_2 , barite and anhydrite are found in the white clouds (white smokers). So-called “snow-blower vents” emit dense clouds of white filaments of native sulphur that is produced from H_2S by sulphur-oxidizing bacteria.

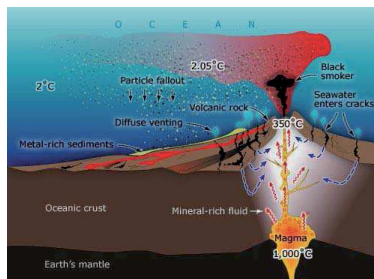


Fig. 34. Black smokers with solutions rich in metals.



Fig. 35. White Smokers.

- **How could Black and White smokers be formed (Fig. 36)?**

1. Cold seawater (2°C) seeps down through cracks into the ocean floor.
2. The seawater continues to seep far in the ocean crust. Energy radiating up from molten rock deep beneath the ocean floor raises the water's temperature to around 350–400°C. As the water heats up, it reacts with the rocks in the ocean crust. These chemical reactions change the water in the following way:
 - i. All oxygen is removed.
 - ii. It becomes acidic.
 - iii. It picks up dissolved metals, including iron, copper and zinc.
 - iv. It picks up hydrogen sulfide.
3. Hot liquids are less dense and therefore more buoyant than cold liquids. So the hot hydrothermal fluids rise up through the ocean crust just as a

hot-air balloon rises into the air. The fluids carry the dissolved metals and hydrogen sulfide with them.

4. The hydrothermal fluids exit the chimney and mix with the cold seawater. The metals carried up in the fluids combine with sulfur to form black minerals called metal sulfides, and give the hydrothermal fluid the appearance of smoke. Many factors trigger this reaction. One factor is the cold temperature, and another is the presence of oxygen in the seawater. Without oxygen, the minerals would never form.

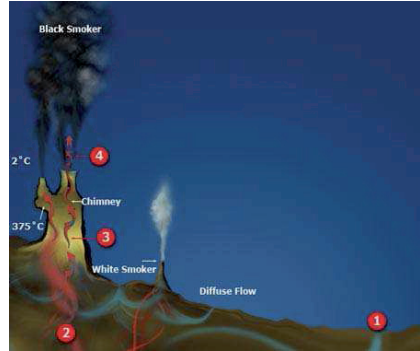


Fig. 36. How are Black and White smokers formed?

In white smokers, the hydrothermal fluids mix with seawater under the seafloor. Therefore, the black minerals form beneath the seafloor before the fluid exits the chimney. Other types of compounds, including silica, remain in the fluid. When the fluid exits the chimney, the silica precipitates out. Another chemical reaction creates a white mineral called anhydrite. Both of these minerals turn the fluids that exit the chimney white.

In other words, the origin of mid-ocean submarine hydrothermal systems is mainly seawater convection in hot young oceanic crust, on top or above the flanks of shallow magma bodies 1 to 3km below the seafloor (Fig. 37). The seawater flows downwards to more than 3km depth through the fractures developed due to the convection current and divergent plate boundaries. At higher temperature and deeper levels, the descending seawater reacts with basalts causing ocean floor greenschist facies metamorphism (Fig. 38). Water oxygen is rapidly consumed by reaction with Fe(II) and new hydrated minerals incorporating OH are formed (e.g. chlorite, amphibole). Consequently, the H^+ increased in the fluid increasing its acidity.

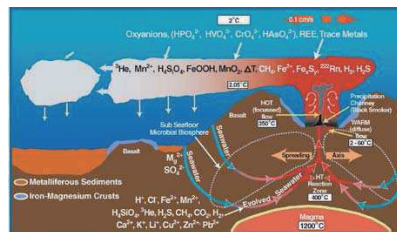


Fig. 37. How are Black and White smokers formed?

Chapter 3: Magmatic Ore Deposits

The acid water dissolves metals and sulphur of the country rocks. Although most of the emitted metals are diluted in ocean water and sediments, approximately 250 metalliferous bodies of economic mass and grade have meanwhile been discovered (Fig. 37).

Beneath vent fields, large Cu, Zn and Au accumulations are probably formed by precipitation because of boiling and vapour loss during de-pressurization (Fig. 39). Metalliferous mud in several depressions of the Red Sea represents the largest known submarine base metal mineralization.

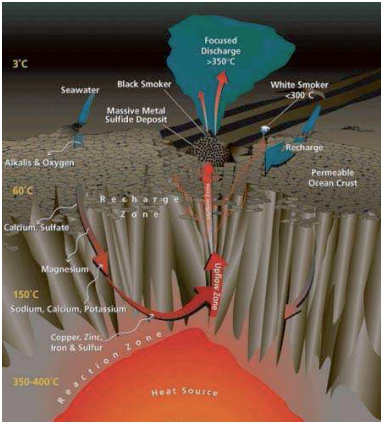


Fig. 38. Formation of White and Black Smokers.

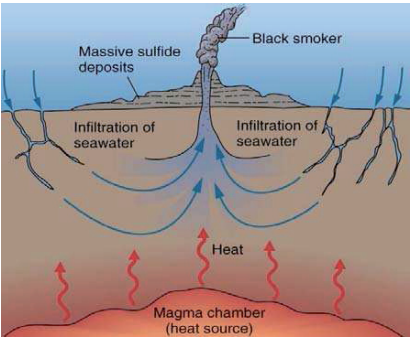


Fig. 39. Young massive sulfide deposits around the smokers.

2.3 Ore formation related to alkaline magmatic rocks, carbonatites and kimberlites

Rocks of alkaline affinity generally have low SiO₂ and high alkali element content, especially of sodium and potassium. They occur mainly in continents, and rarely within oceanic plates. An anorogenic setting is affirmed by the existence of these rocks near continental rifts, over heat anomalies of the mantle (hot spots, plumes, superplumes). The alkaline magmas originate by a low degree of partial melting of enriched mantle material may stem from subducted oceanic crust, or more probably, from metasomatized lithospheric mantle.

Nephelinite (alkaline) magma is the most common mafic alkaline liquid that crystallizes to give a range of igneous rocks (termed the ijolite suite) (Table 1). They are typically associated with the much rarer carbonatites that have a more prominent metallogenetic role.

Table 1. Nomenclature of some alkaline igneous rocks.

Basante	feldspathoid-bearing basalt. Usually contains nepheline, but may have leucite + olivine
Tephrite	olivine-free basante
Leucite	a volcanic rock that contains leucite + clinopyroxene ± olivine. It typically lacks feldspar
Nephelinite	a volcanic rock that contains nepheline + clinopyroxene ± olivine. It typically lacks feldspar. Fig. 14-2
Urite	plutonic nepheline-pyroxene (aegirine-augite) rock with over 70% nepheline and no feldspar
Ijolite	plutonic nepheline-pyroxene rock with 30-70% nepheline
Melilitite	a predominantly melilitite - clinopyroxene volcanic (if > 10% olivine they are called olivine melilitites)
Shoshonite	K-rich basalt with K-feldspar ± leucite
Phonolite	felsic alkaline volcanic with alkali feldspar + nepheline. See Fig. 14-2. (plutonic = nepheline syenite)
Comendite	peralkaline rhyolite with molar (Na ₂ O+K ₂ O)/Al ₂ O ₃ slightly > 1. May contain Na-pyroxene or amphibole
Pantellerite	peralkaline rhyolite with molar (Na ₂ O+K ₂ O)/Al ₂ O ₃ = 1.6 - 1.8. Contains Na-pyroxene or amphibole
Lamproite	a group of peralkaline, volatile-rich, ultrapotassic, volcanic to hypabyssal rocks. The mineralogy is variable, but most contain phenocrysts of olivine + phlogopite ± leucite ± K-richite ± clinopyroxene ± sanidine. Table 19-6
Lamprophyre	a diverse group of dark, porphyritic, mafic to ultramafic hypabyssal (or occasionally volcanic), commonly highly potassic (K>Al) rocks. They are normally rich in alkalis, volatiles, Sr, Ba and Ti, with biotite-phlogopite and/or amphibole phenocrysts. They typically occur as shallow dikes, sills, plugs, or stocks. Table 19-7
Kimberlite	a complex group of hybrid volatile-rich (dominantly CO ₂), potassic, ultramafic rocks with a fine-grained matrix and macrocrysts of olivine and several of the following: ilmenite, garnet, diopside, phlogopite, enstatite, chromite. Xenocrysts and xenoliths are also common
Group I kimberlite is typically CO ₂ -rich and less potassic than Group 2 kimberlite	
Group II kimberlite (orangeite) is typically H ₂ O-rich and has a mica-rich matrix (also with calcite, diopside, apatite)	
Carbonatite	an igneous rock composed principally of carbonate (most commonly calcite, ankerite, and/or dolomite), and often with any of clinopyroxene alkalic amphibole, biotite, apatite, and magnetite. The Ca-Mg-rich carbonatites are technically not alkaline, but are commonly associated with, and thus included with, the alkaline rocks. Table 19-3

For more details, see Sørensen (1974), Streckisen (1978), and Woolley *et al.* (1996)

“Shallow” carbonatitic and deep kimberlitic melts with high CO₂ and low H₂O content originate in lithospheric mantle at 120–260km depth. The high gas content facilitates rapid rise of magma diapirs to the surface where eruption takes place (Fig. 40).

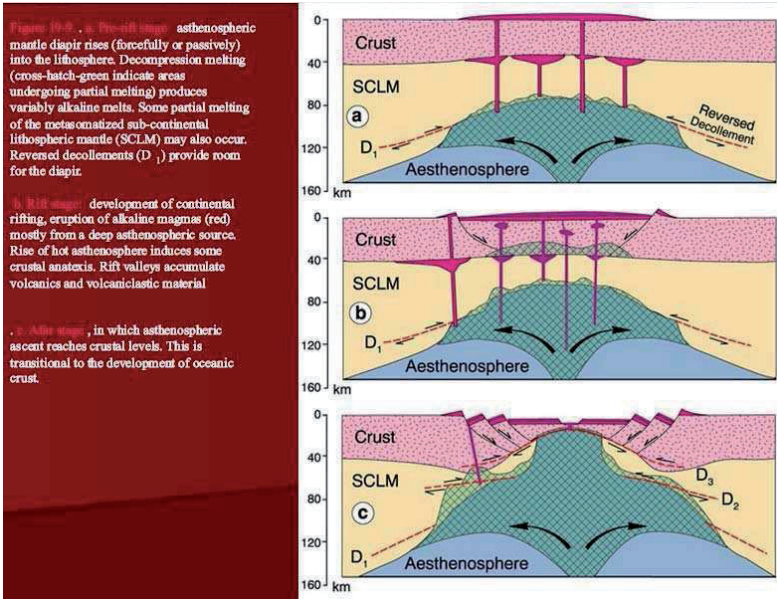


Fig. 40. Origin of alkaline igneous rocks.

There are two hypotheses about the origin of alkaline rocks and carbonatites (ARCS) (Fig. 41). In the plume model, ARCs are derived from mantle plumes (here defined simply as magma sources of distinctive chemical composition within the convecting mantle). In the deformed alkaline rocks and carbonatites (DARC) model, ARCs are derived from melting that involves deformed alkaline rock and carbonatite material that was carried into the lithospheric mantle during an ancient subduction episode.

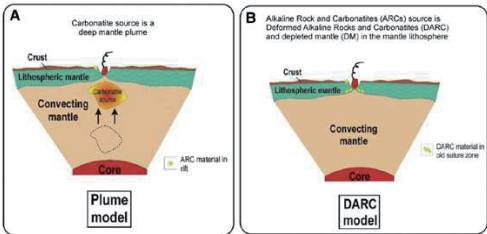


Fig. 41. Plume and DARC model for the origin of alkaline igneous rocks.

2.3.1 Carbonatites

Carbonatites are igneous rocks with more than 50% of carbonate minerals. They are further subdivided depending on the nature of the carbonates

(calcite, dolomite, and ankerite) and the silicate phases (biotite, pyroxene, amphibole, etc.).

There are three possible models for the generation of carbonatitic magmas (Fig. 42):

- i. direct partial melting of the upper mantle peridotite induced by addition of CO₂,
- ii. fractional crystallization of a nepheline normative, silica-undersaturated, relatively alkali rich silicate magma containing dissolved CO₂ and probably also H₂O; and
- iii. separation of an immiscible carbonatite melt from an alkali-rich or Ca-rich silicate magma.

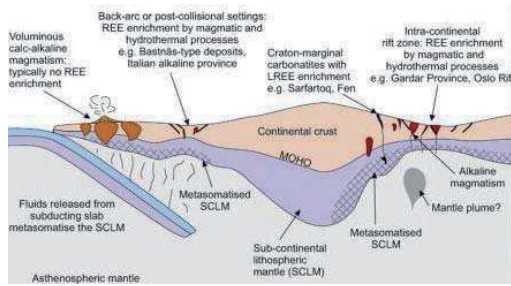


Fig. 42. Tectonic emplacement of Carbonatites.

Field relations do not support the fractional crystallization model either, because carbonatites are not found associated with a differentiated series of silicate rocks. The liquid immiscibility model, on the other hand, is supported by several lines of field and chemical evidence.

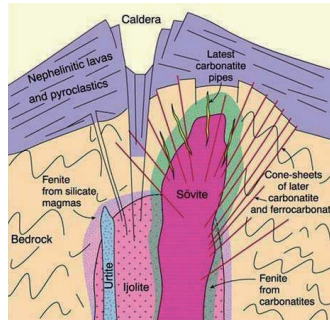


Fig. 43. Carbonatites.

Carbonatites occur as both intrusive and extrusive bodies - the former as plutonic and hypabyssal dikes, sills, sheets, pipes, stocks, and more irregular bodies; the latter as flows and pyroclastics (Fig. 43).

Anomalous amounts of rare earth elements (REE) are remarkable features of carbonatites, especially of the light REE Elements (lanthanum to samarium), P, F, Th, Ti, Ba, Sr, and Zr. Half of all known carbonatites occur along the East African Rift System.

Metals exploited from complex intrusions of carbonatite, alkali-pyroxenites and nepheline syenites include:

- i. Metallic, such as copper, rare earth elements, iron-titanium-vanadium, uranium-thorium and zirconium;
- ii. Non-metallic, such as vermiculite, apatite, fluorite and barite, and limestone.

Nepheline syenite is a good source for Al in ceramics industry.

The most important mineral products of carbonatites probably are calcite for cement and apatite for phosphatic fertilizer. Many carbonatites contain traces of Th-bearing monazite, pyrochlore, and uranothorianite, which are useful for outlining carbonatite bodies by radiometric surveys. The principal metals for which the carbonatites are considered a major resource are niobium and REE; some carbonatites also contain significant concentrations of Fe (magnetite, hematite), Ti (rutile, brookite, ilmenite, perovskite), Cu sulfides, barite, fluorite, and strontianite, which may be recoverable as byproducts. Pyrochlore ($\text{CaNaNb}_2\text{O}_6\text{F}$) is by far the most abundant primary niobium mineral in carbonatite associations and it is found in nearly all rock types of carbonatite complexes in accessory amounts.

2.3.2 Kimberlites

Kimberlites are derived from the Earth's mantle at more than 140km depth (Figs. 44 and 45). They are petrographically variable rocks comprise strongly altered breccias and tuffs. Basically, Kimberlites are porphyric, SiO_2 undersaturated, K-rich (1–3 wt.% K_2O) peridotites with xenoliths, and xenocrysts of diamond

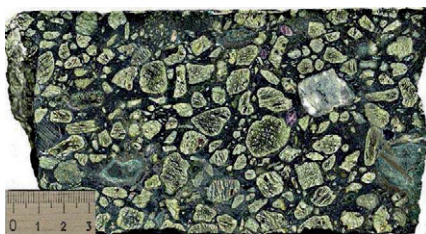


Fig. 44. Kimberlites.

and olivine in a carbonated and serpentinized groundmass. Kimberlite is a *hybrid* rock, which does not consider a true representation of melt composition. Kimberlite is the rock which contains diamond.

- **Diamond**

Diamonds are formed under hot and high pressure conditions. The physical and chemical conditions where diamonds form only exist in the mantle. In the

upper mantle, diamonds may be a common mineral! Diamond is associated with volcanic features called diatremes. A diatreme is a long, vertical pipe formed when gas-filled magma forces its way through the crust to explosively erupt at the surface (Figs. 45 and 46).

Kimberlite is associated with some diatremes that sometimes contain diamonds. Diamonds are xenoliths carried up from deep sources in the mantle, and often occur in association with other gem minerals including garnet, spinel and diopside inside the kimberlite. They are most extensively mined from Kimberlite pipes or from alluvial gravels derived downstream from diamond source areas.

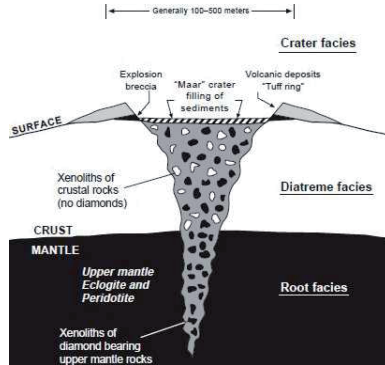


Fig. 45. Kimberlites in diatremes.

Whenever carbon occurs as a free species, diamonds have the potential to form (Fig. 47). Diamonds are stable under the high pressure and temperature conditions that are only met at great depth in the earth's mantle. Continental regions that long ago ceased participating in active plate tectonic processes such as rifting, mountain building, or subduction are known as continental cratons and has the Archean age. Diamonds always occur within the Cratons, especially those hosted in Kimberlite, the main carrier and hence “ore” of gem-quality diamond. Withering of Kimberlite, releases the diamonds to the regolith (Fig. 46). When transported by rivers, the alluvial diamonds are concentrated in the placer.

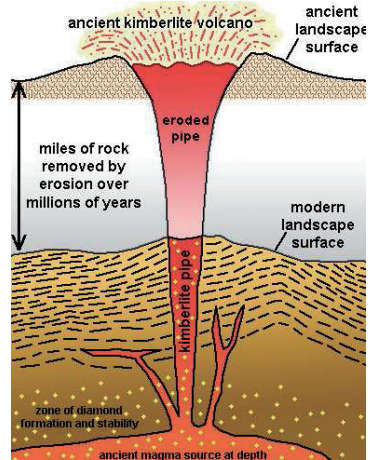


Fig. 46. Diamond-bearing Kimberlite pipes are diatremes that originate in the mantle.

In tectonically stable areas, the mantle keel under each craton is at high enough pressure and comparatively low temperature to allow diamonds to crystallize whenever they receive fluids saturated in carbon from the

underlying convecting mantle (Fig. 48). The keel bottom can be viewed as an “ice box” to store diamonds and keep them from entering mantle circulation, to be sampled by a rising Kimberlite magma (the Phenocryst model). The Kimberlite eruptions that transport diamonds to the surface also carry samples of lithospheric mantle rocks called xenoliths. Both peridotite and eclogite contain diamonds, but intact peridotites subducted to the surface – ophiolites - with their diamonds are rare, while eclogites (high pressure metamorphosed basalt/gabbro) with their diamonds in place are common.

In tectonically un-stable environment, diamond is destroyed in the volcanism, mountain-building, and intrusive magmatism near the earth’s surface, where pressures, temperatures, and oxidizing conditions are not suitable for diamond to crystallize or remain stable. However, diamonds can be found in non-kimberlitic rocks formed in tectonic areas that were once active (Fig. 49). Subduction-related (non-kimberlitic) magma type can carry diamonds from the mantle. Late-stage subduction-related magma can produce a rock called a lamprophyre and lamproite as dikes carrying diamonds.

Diamonds are known to be carried to the earth’s surface in only three rare types of magmas (Table 2): kimberlite, lamproite, and lamprophyre. Of the three types, kimberlites are by far the most important, with several hundred diamondiferous kimberlites known. In

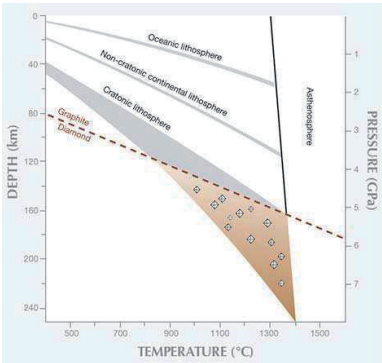


Fig. 47. The stability fields of graphite and diamond in relation to the convecting mantle (asthenosphere) and the lithospheric mantle. Note that only the cratonic lithospheric keel is cold enough at high enough pressures to retain diamonds.

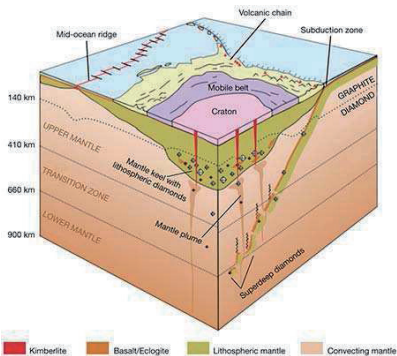


Fig. 48. The relationship between a continental craton, its lithospheric mantle keel (the thick portion of the lithospheric mantle under the craton), and diamond stability regions in the keel and the convecting mantle. Under the right conditions of low oxidation, diamonds can form in the convecting mantle, the subducting slab, and the mantle keel.

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general, all three magma types are: (1) derived by small amounts of melting deep within the mantle; (2) relatively high in volatile (H_2O , CO_2 , F, or Cl) contents; (3) MgO-rich; (4) marked by rapid eruption; and (5) less oxidizing than more common basaltic magma.

Economically important kimberlites appear to be localized in regions underlain by portions of the cratons which are older than 2.4 Ga. These include the diamond-bearing kimberlites of Africa (Angola, Botswana, Lesotho, Sierra Leone, South Africa, Swaziland, Tanzania), Russia (Yakutia), Australia (Western Australia), and the recently discovered kimberlite pipes in Canada (NWT). Some kimberlites are non-diamondiferous either because the magma was generated outside the P-T stability field of diamond or because the magma never picked up any diamond xenocryst due to the non-uniform distribution of diamonds in the upper mantle.

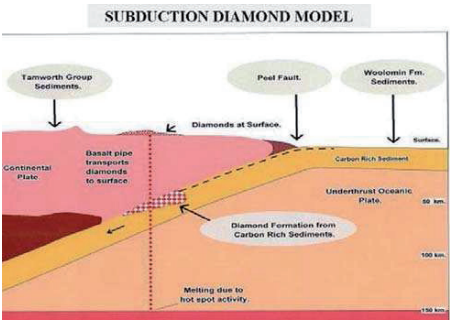


Fig. 49. Diamond in non-kimberlitic rocks.

Table 2. Characteristics of diamond-carrying magmas.

Characteristic	Kimberlite	Lamproite	Lamprophyre
Color of rock in outcrop	Green, dark bluish green	Dark gray, black	Dark gray, black
Volatiles	$CO_2 > H_2O$, halogens (Cl, F)	$H_2O > CO_2$, halogens	$H_2O > CO_2$, halogens
Composition	Not peralkaline, hybrid, $K_2O > Na_2O$	Peralkaline, hybrid to magmatic	Peralkaline, magmatic
Setting	Cratonic	Mobile belt, craton margin	Subduction zone
Depth of origin	200–300 km mantle, can be sourced >400–700 km	>140 km, but probably not much deeper	>140 km, but probably not much deeper
Eruptive style	Explosive volcanic pipes, pyroclastic	Small volcanic pipes and cones	Dikes
Diagnostic phenocryst minerals	Olivine	Ti-phlogopite	Biotite, amphibole
Typical matrix mineralogy	Olivine, carbonate	Phlogopite	Feldspar
Diamond potential	Common, can range to high diamond grade	Rare, but can range to high diamond grade	Rare, few diamondiferous known

2.4 Granitoids and ore formation processes

Granitoids are felsic plutonic rocks with more than 20 % quartz. The ore formation potential depends on origin and evolution of the parental granitoid.

There are many variables to control the fertility of acidic magma, these are:

1. the plate tectonic setting,
2. the nature of source rocks,
3. P/T-parameters of melting,
4. content of water and other volatiles,
5. the depth of intrusion,
6. coeval tectonic deformation,
7. partial pressure of oxygen (redox state) of the melt,
8. assimilation of country rocks and the evolution of the magma by fractionation,
9. cooling and crystallization including fluid segregation.

Trace elements and isotope systems in granitoids provide valuable information on the source rocks of granitoids. Fundamentally distinct sources of granitoids are:

1. Peridotites of the Earth's upper mantle (asthenosphere, lithosphere). M-type granitoids are sourced in the mantle. They intrude the crustal rocks of ophiolites in the form of plagiogranite and quartz diorite, and the thick volcanic piles of primitive oceanic island arcs. Typical ore deposits associated with M- type granitoids are copper-gold porphyries and hydrothermal gold.
2. Magmatic and metamorphic rocks of the deep continental crust (infracrustal). I-type granitoids originate by melting of pre-existing infracrustal igneous rocks. I-type granitoids are the most common intrusive magmatic rocks. They display an abundance of hornblende and higher concentrations of Ca, Na and Sr compared with granites derived from sediments. Examples of I- type granitoids are tonalites and granodiorites. The magma formed the I-type granitoids are undersaturated with water, which enabled them to rise to the surface, forming volcanic rocks (e.g. andesite and dacite). Accessory minerals of I-type granitoids are often magnetite and titanite (magnetite-series magmatic rocks). This is due to a commonly higher oxidation degree of

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I-type magmas, although reduced I-type granitoids are known. Characteristic ore deposits related to I-type granitoids are:

- i. the iron oxide-copper-gold (U-REE) deposits (IOCG),
 - ii. copper-molybdenum porphyries,
 - iii. Mo-W Cu skarn,
 - iv. hydrothermal lead-zinc and
 - v. certain gold and silver ores.
3. Clastic metasediments and metamorphic equivalents (supracrustal). S-type granitoids originate by continental collision and deep subduction of sediments to great depths and high temperatures. Resulted melts are mainly leucocratic, SiO₂ rich rocks of a monzogranitic nature, often with muscovite and biotite. Accessory minerals include cordierite, garnet, kyanite and ilmenite (ilmenite-series magmatic rocks). The oxidation of these magmas is low, due to organic carbon in the source sediments. The water of the melts is derived by dehydration of muscovite in the metasediments. Highly fractionated intrusions could have the following ores: tin, tungsten and tantalum ore deposits.
4. Restites of sediments and of magmatic rocks that have experienced earlier anatexis before a later melting event. A-type granitoids “abnormal, anhydrous, alkali rich, aluminous and anorogenic” are the product of repeated melt-extraction from the same source rocks. Some granites that have A-type characteristics may be derived by extreme fractionation of I- and S-type magmas. With every cycle of melting the source rocks acquire a more pronounced restite composition, marked by enrichment of less mobile substances. Another possible source of A-type magma is lithospheric mantle and not all A-granites are anorogenic. Typical A-type granites are the alkali granites of continental Rifts. Volcanic equivalents include tin-rich topaz rhyolites in fields of crustal distension. Two different ore associations occur with A-type granitoids:
- i. Sodium-rich granites, contain concentrations of niobium, uranium, thorium, rare earth elements and some tin.
 - ii. Potassium- rich granites with profuse hydrothermal silicification, tourmalinization and acidity produce deposits of tin, tungsten, lead, zinc and fluor spar. This association may occur within the

granite body (endogranitic greisen, pegmatite, and porphyry stockwork ore) or in vein fields within intruded rocks (exogranitic).

Specialized and mineralized (parental) granites are distinguished by geochemically elevated content of metals, such as Sn, W, Nb, Ta, Mo, U, Th, REE, Rb, Cs, Li, Be, often F (the latter include "topaz granites") and P. In specialized granites, rare elements are enriched in silicates and accessory minerals. Mineralized or parental granites, in contrast, stand out by their close relations to ore-grade concentration of rare elements.

There are some trace elements found in granitic melt. These elements because of their sizes, charges and/or chemical affinity are strongly partitioned into the fluid phase and not to be crystallized with granitic silicate minerals. These elements are called "granophile trace elements". Eventually, they are concentrated to form mineral deposits related to granitic intrusions. These deposits are also called granophile mineral deposits. B and Be which has very small sizes that can not substitute in the lattices of normal silicate minerals of granites. So, they form ore minerals such as tourmaline and beryl.

The geochemical changes from ordinary to mineralized granitoids are mainly caused by a process system, which is generally termed "magmatic fractionation". Granites with extreme chemical fractionation are the source (and often the hosts) of deposits of rare elements including Sn, Li and Be. They are enriched in large ion lithophile elements (LILE) such as Rb and Cs, and high field strength elements (HFSE) such as P, Y, Zr, Hf, Nb, Ta, Th and U. The increase of the magmatic differentiation shows how tantalum (an incompatible element) is continuously enriched by increasing differentiation of successively more fractionated granite melt and finally reaches exploitable grades. The increasing differentiation of magmas is caused by fractional crystallization, early crystal settling and/or removal of liquid melt.

- **Fertility of granitic magma**

Fertile granitoid magmas are distinguished by high content of volatiles. Volatiles collect the rare elements that form ore. In addition, lower density, viscosity and solidus temperature of a melt increases its mobility. Low magmatic temperatures and high salt concentrations favour the fractionation of metals into the fluid phase. Oxygen fugacity (pressure) in the melt is an

important control. High oxygen in granitic magma (magnetite series) causes depletion of tin (Sn) and tungsten (W) in the liquid and in late fluids (let them devoid of), because these metals are abstracted – taken - in dispersed accessory minerals already during main-stage crystallization. Behaviour of copper (Cu) and uranium (U) is quite the reverse (compared with that of Sn and W). Oxidized magmas dissolve more sulphur, as an “anhydrite component” and the derived fluids may produce large Cu-Au deposits, i.e., mineralization occurs within the granite during crystallization. In “reduced” granitic magmas (ilmenite series), early sulphur saturation causes formation of dispersed sulphide droplets that collect copper and gold.

2.5 Ore deposits in pegmatites

Pegmatites crystallize from highly fractionated hydrous residual melt batches of felsic magma bodies that are enriched in volatiles and incompatible trace elements (Fig. 50). Pegmatites are characterized by:

- i. coarsely crystalline textures,
- ii. occasionally by giant crystals,
- iii. miarolitic cavities,
- iv. minerals of rare elements.

Most pegmatites are related to granites and have a paragenesis of orthoclase (perthite), microcline, albite, mica and quartz. Common minor minerals include tourmaline, topaz, beryl, cassiterite and lithium minerals. Felsic pegmatite melts intruding ultramafic rocks suffer desilication resulting in plumasites that are characterized by corundum, kyanite and anorthite.

Gabbro pegmatites are derived from mafic magmas and are composed of anorthite, pyroxene, amphibole, biotite and titanomagnetite, occasionally including carbonates and sulphides (Fig. 51). Iron-rich ultramafic pegmatites composed of olivine. Rare syenite



Fig. 50. Granitic Pegmatites.



Fig. 51. Gabbro Pegmatites.

pegmatites with microcline, nepheline, apatite, niobium and rare earth element minerals are related to alkaline intrusions (Fig. 52).

Granitic pegmatites occur in the form of dykes, oval and lenticular bodies. Most pegmatite bodies are relatively small with tens of metres thickness and a length of a few hundred metres. Some pegmatites occur at the roof of granite and form a thin shell between the intrusion and the roof rock. Granitic pegmatites may be isotropic (homogeneous) (without a change of mineralogy or texture from wall to wall) OR anisotropic (inhomogeneous - “zoned” or “complex” pegmatites) (Fig. 53).



Fig. 52. Sylvanite Pegmatites.

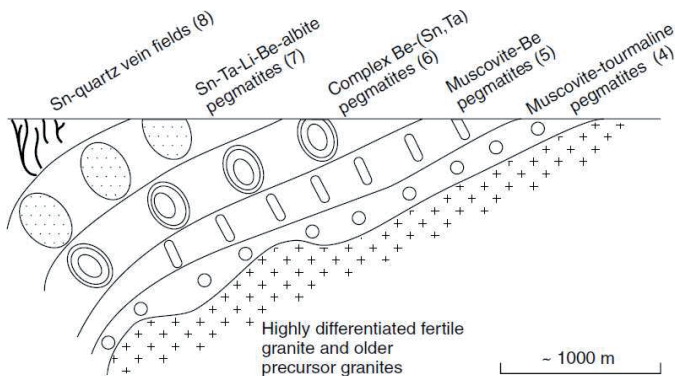


Fig. 53. External zonation of rare element pegmatites and cassiterite quartz veins near fertile granites in Central Africa.

The internal zonation of complex pegmatites reflects crystallization from the walls to the centre of a pegmatite body (Fig. 54).

The following zones are distinguished:

1. Border zone: often fine-grained = aplitic, and very thin;
2. Wall zone: coarsely crystalline

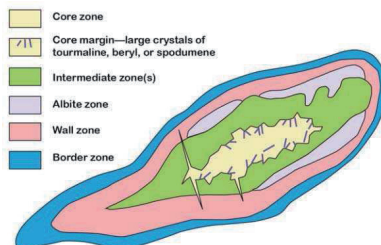


Fig. 54. Deposit-scale zoning patterns in an idealized pegmatite.

- with exploitable muscovite and beryl;
3. Intermediate zones: albititic with microcline and contain the valuable minerals (cassiterite, columbite, spodumene, beryl, etc.);
 4. Core: which is commonly a solid mass of barren grey or white quartz, but may also contain feldspar, tourmaline and spodumene.

The internal zonation in complex pegmatites might have two causes: i) fractional crystallization in a closed system; or ii) repeated injection of new melt batches in an open system. A chemical exchange directed from enclosing rocks to the pegmatite is possible. The wall zone could contain tourmaline-rich due to reaction of iron and magnesium mobilized from the host rocks with boron from the volatile phase of the pegmatite.

The pegmatite melts are ejected along with enrichment of water, B, F, P, Sn, Rb and other incompatible elements, while the main magma body crystallizes. Another possibility is that small pegmatitic melt batches rise directly from the source region of the “parent” granite.

Pegmatites may host many useful raw materials. These include ores of Be, Li, Rb, Cs, Ta > Nb, U, Th, REE, Mo, Bi, Sn and W, the industrial minerals muscovite, feldspar, kaolin, quartz, spodumene, petalite and fluorite, and gemstones as well as rare mineral specimens (emerald, topaz, tourmaline, ruby, etc.). The derivation of pegmatites from I-, S- and A-type granites is probably the main control of the availability of specific elements for enrichment.

2.6 Hydrothermal ore formation

The term “hydrothermal water” applies to subsurface water with a temperature that makes it an agent of geological processes (Fig. 55), including hydrothermal ore formation. “Geothermal water” is a subgroup of hydrothermal solutions that

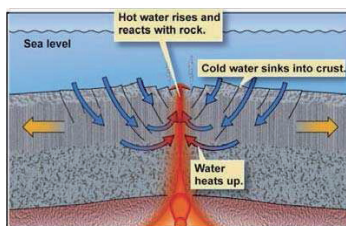


Fig. 55. Hydrothermal water.

occur near the Earth's surface and is mainly used as an geothermal energy source.

Thermal springs are common indicators of geothermal reservoirs at depth.

Many hot springs and geysers currently display precipitation of minerals and ore. Isotopic investigations revealed that many geothermal and hydrothermal waters are not of magmatic but of meteoric derivation (i.e. from local precipitation) (Fig. 56). Hydrothermal activity in undersea volcanoes is largely the result of sea water descending into the crust, being heated up and then chemically breaking down the surrounding rocks as it rises back up to the sea bed. These mineral rich fluids then re-enter the water column either diffusely over a wide area, or out of one of many vents in a hydrothermal field. Most hot waters are dilute solutions of chloride, carbonate and sulphate, but dissolved silica, boron and sulphide are also common.

- **Hydrothermal convection**

Hydrothermal convection cells are established where heat sources below the surface coincide with permeable flow paths, often provided by extensional tectonic deformation (Fig. 57). Cold infiltrating surface and groundwater is drawn to the “heat exchanger” at depth. The lower density of hot compared to cold water causes ascent of hydrothermal solutions and establishes hydrothermal convection.

- **Chemical and physical characteristics of hydrothermal solutions**

The chemistry of hydrothermal solutions is variable and a result of interaction between rocks and hot water. Factors like initial state of rock and water, the water/rock mass ratio, temperature, chloride concentration, pressure and redox state control the chemistry of hydrothermal solutions. The fraction of dissolved matter in hydrothermal solutions varies from less than 1 to over 50 wt.%. Chlorine and sulphur are the most important anions. Salinity ranges from very low to more than 50% and the source of salinity (e.g. halite dissolution, evaporation of seawater, etc.) is detectable by determination of halogens and electrolytes.

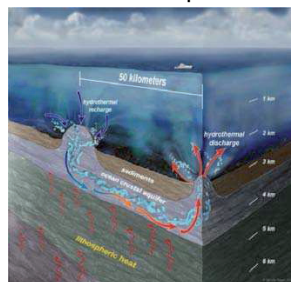


Fig. 56. Hydrothermal water.

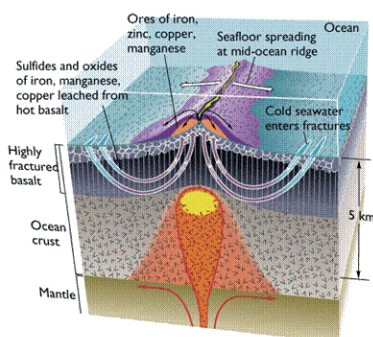


Fig. 57. Hydrothermal water.

Metals are to some extent dissolved as simple ions or ion pairs, but more commonly in the form of complex ions, which combine chlorine, dissociated OH groups and bisulphides, as well as NH_3 , H_2S and CO_3 . Metal concentrations range from less than 1 to several 1000 ppm (parts per million, equal to gram/tonne).

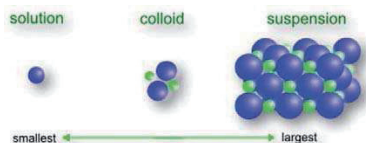


Fig. 58. Colloids.

Even higher concentrations in solution are possible when metals are part of complex ions. Hydrothermal solutions carry metals not only in dissolved form but also as colloidal particles. Colloids are tiny particles (1–1000 nm), which are quite common in many natural waters, usually at low concentrations (Fig. 58). High concentrations of dispersed colloids in water are called hydrosols. In many cases, hydrosols are the precursors of gels. Hydrosols and gels may form by local supersaturation of a substance, because of a sudden change of pH, T, P or Eh.

A fluid comprising CO_2 or CH_4 in addition to water has a high carrying capacity that depends on pressure and density variations. Very small variations cause either dissolution or precipitation of solids. Hydrogen ion activity (pH) of hydrothermal solutions varies from moderately acidic to moderately Alkalic (exceptions could occur, acidic conditions, for example, cause formation of kaolinite, alunite or topaz from feldspar). Deep hydrothermal water is normally reduced; oxygen content may increase near the surface by mixing with fresh meteoric water (Fig. 59). Bituminous substances are a common accessory in hydrothermal deposits. This can be a sign that the hydrothermal solutions were sourced in basinal sediments (e.g. diagenetic formation water mixed with hydrocarbon fluids).

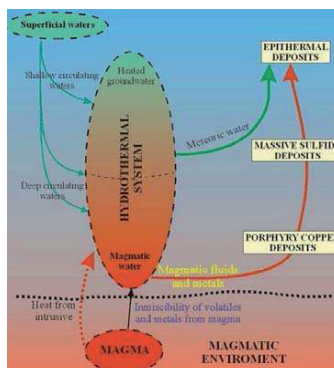


Fig. 59. Magmatic, metamorphic and groundwater fluids may interact in hydrothermal systems to different degrees.

Possible phase states of hydrothermal waters are liquid, gaseous (vapour) and fluid (fluid= supercritical “gas” or “liquid”). Many hydrothermal deposits were formed by supercritical fluids (water reaches its supercritical state at $T >$

374 °C and $P > 225$ bar (Fig. 60)) (increasing salinity moves the critical point to higher T and P). Similar to gas, supercritical fluids have a smaller viscosity, higher diffusivity and mobility (a supercritical fluid is any substance at a temperature and pressure above its critical point, where distinct liquid and gas phases do not exist. It can effuse through solids like a gas, and dissolve materials like a liquid. In addition, close to the critical point, small changes in pressure or temperature result in large changes in density).

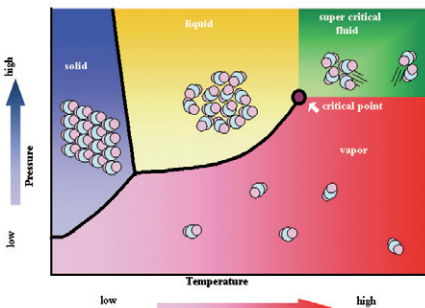


Fig. 60. Supercritical conditions of water.

- **How could ore and gangue minerals precipitate from hydrothermal solutions?**

1. Decreasing temperature and pressure reduces solubility of metals in hydrothermal solutions. Precipitation is a function of the relative stability of metal complexes and decreasing temperature often results in the common sulphide precipitation sequence from early Cu to Zn, Pb, Ag and finally Hg.
2. Pressure drops may cause fluid immiscibility, such as the formation of two fluids (e.g. aqueous and carbonic) from an originally homogeneous fluid (aqueous-carbonic). Pressure drop can change pH, fO_2 and temperature, thus inducing mineral deposition. Rapid pressure fluctuations are typically caused by tectonic events.
3. Falling pressure associated with boiling changes chemical properties of a hydrothermal solution (concentration, pH, Eh, stability of complex ions), which consequently reduces the solubility of dissolved matter (Fig. 61). The term “effervescence” is preferably used in place of “boiling” when gas bubbles form that are not vapour of the host liquid (e.g. carbon dioxide gas bubbles in water). Yet like boiling, effervescence may also induce rapid precipitation of minerals.
4. Mixing of chemically different waters induces deposition of ores and minerals. A common example is the formation of barite. Barite ($BaSO_4$)



Fig. 61. Effervescence of water.

is precipitated when ascending chloride solutions with dissolved barium ions encounter sulphate-ion bearing water (e.g. seawater).

5. The reaction of hydrothermal solutions with host rocks or with previously deposited ore minerals is a very efficient means of immobilizing dissolved elements (Fig. 62). When metal-bearing solutions encounter sulphide minerals, the more noble metals are precipitated, whereas the less noble elements pass into solution:

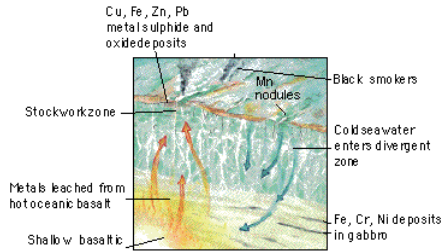
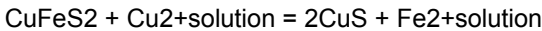


Fig. 62. Circulation of fluids and precipitation of mineral deposits.



This selective precipitation of more noble metals from solution by exchange with less valuable elements is a function of electronegativity, ionization potential, electron affinity, redox potential and the energy of chemical bond formation.

6. Host rocks exert a strong control on noble metal enrichment. Deposition of gold is explained by reaction of sulphide solutions with reduced iron of doleritic host rocks, forming pyrite "sulphidation". When sulphidation happens, a radical decrease of reduced sulphur in the hydrothermal solutions occurs causing gold precipitation (Fig. 63).



Fig. 63. Gold.

7. Organic substances (coal, kerogen, oil, gas) also motivate immobilization of many metals by adsorption or reduction (Fig. 64). Gold ore veins and the metasomatic gold orebodies of Carlin, USA are enriched where host rocks contain kerogen-rich layers. Sulphide precipitation in Mississippi Valley deposits is often caused by reaction between solutions and the organic substance

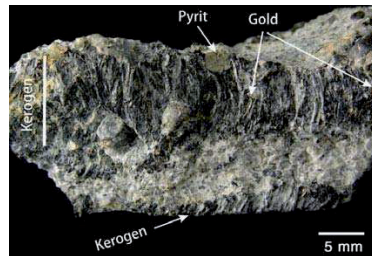


Fig. 64. Carbonaceous layer of sediment rocks, essentially consisting of kerogen, gold and pyrite .

of host rock carbonates.

8. Incompletely oxidized sulphur (e.g. thiosulphate, $\text{S}_2\text{O}_3^{2-}$, polysulphides, colloidal sulphur) supports high metal content in solution. These compounds, however, are easily reduced by contact with organic matter so that metals are instantly immobilized as sulphides. An indirect consequence is the precipitation of gangue, such as barite and fluorite.
9. Although reduction is a frequent means of ore mineral deposition, oxidation can have a similar role, most often concerning iron and manganese. Hydrothermal solutions transport these metals in reduced form (Fe^{2+} , Mn^{2+}) and precipitation of haematite (Fig. 65), magnetite or pyrolusite requires oxidation to Fe^{3+} or Mn^{4+} .
10. Contact of hydrothermal metal-bearing solutions with carbonate rocks is a frequent factor of ore precipitation. Individual agents include:
 - i. the “pH shock” upon contact with alkaline rocks (carbonates) and formation fluids;
 - ii. a larger permeability compared with pelitic country rocks;
 - iii. a higher solubility of carbonates in acidic or CO_2 -rich solutions (which may result in the formation of “hydrothermal karst”) and
 - iv. mixing with formation water in carbonate rocks.

Orebodies in carbonates take the form of veins, breccia, karst pipes and stratiform orebodies with irregular outlines (“mantos”). When the replacing masses consist of sulphides, dissolution of the original carbonate rock and replacement (“metasomatism”) by ore take place. “Metasomatism” is used for cases where only cations are exchanged (e.g. siderite in limestone).



Fig. 65. Haematite ore.



Fig. 66. stratiform Zn-Pb-Ag-rich, generally Fe and Cu sulfide-poor, massive and semi-massive sulphide.

- **Source and origin of hydrothermal fluids and solutions**

Source and origin of hydrothermal fluids and solutions may be related to quite different geological process systems:

- i. magmatism (exsolution of an aqueous fluid phase from silicate magma);
- ii. heating of meteoric, oceanic or formation water by convection within or near cooling intrusions, including large faults or uplifted hot metamorphic complexes;
- iii. diagenesis (mainly physical dehydration of sediments by increasing pressure and temperature because of increasing overburden, thrust sheet superposition, or accretion on active continental margins);
- iv. metamorphism (mainly chemical dehydration of minerals that include OH-groups in their crystal lattice, caused by prograde metamorphic reactions);
- v. mixing of two or more of the mentioned source systems.

- **Diversity of hydrothermal ore deposit**

In the Earth's crust, the hydrothermal ore deposits occur in a fascinating diversity (Fig. 67):

- i. Veins;
- ii. metasomatic bodies in carbonates;
- iii. breccia ores in magmatic rocks ("porphyry deposits");
- iv. ore stockworks and pipes;
- v. volcanogenic terrestrial and submarine exhalations;
- vi. stratiform base metal ore beds in marine sediments (sedimentary-exhalative ore SEDEX) and
- vii. stratabound diagenetic Pb-Zn-Ba-F deposits in marine carbonates.

- **Classification of hydrothermal solutions/ore deposits**

Traditionally, hydrothermal ore deposits were grouped according to assumed formation temperatures into (Table 3 and Fig. 68):

1. hypo- or katathermal (500–300 °C);
2. mesothermal (300–200 °C) and
3. epithermal (below 200 °C).

This classification was quietly abandoned because temperatures vary widely even within one single hydrothermal system. The above terms are still used in a very wide sense, indicating rather depth than temperature.

Clearly, depth (or pressure) is a much more useful criterion to describe related groups of hydrothermal deposits. Therefore, the terms “epi-, meso- and hypozonal” similar to the notations referring to metamorphism or the intrusion depths of granites.

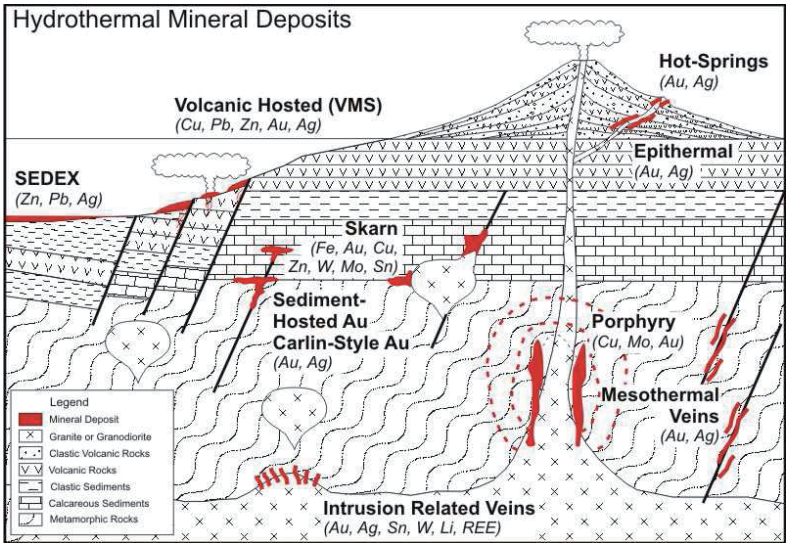


Fig. 67. Hydrothermal mineral deposits, are formed by a process involving the dissolution, transportation, and precipitation of metals in “hot” hydrothermal fluids. These deposits can form at or near the earth’s surface or they can form deep in the crust and show distinct characteristics based on the depth of formation. Each mineral deposit shows distinct characteristics which are controlled by the characteristics of the mineralizing fluids, the characteristics of the host rocks and the solubility of the elements of interest.

Table 3. Depth-zone classification of hydrothermal mineral deposits.

Epizonal	150–300°C	0.5–1.5 kbar	<6 km depth
Mesozonal	300–475°C	1.5–3.5 kbar	6–12 km
Hypozonal	475–700°C	3–6 kbar	>12 km

Chapter 3: Magmatic Ore Deposits

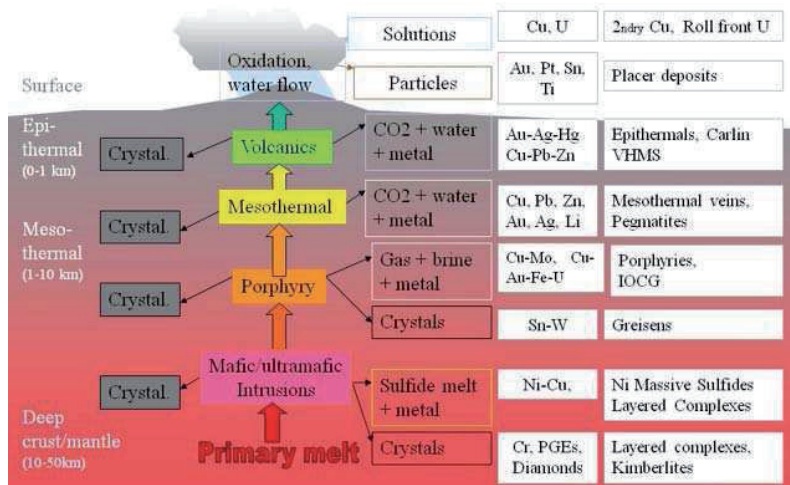


Fig. 68. Types of hydrothermal ore deposits.

2.7 Skarn- and contact-metasomatic ore deposits

Many ore deposits are formed close to intrusive igneous rock bodies. The location of the ore may be at the immediate contact between the intrusion and the host rocks, or at a certain distance (Fig. 69). In the first case, the host rocks will be affected by contact metamorphism due to heating (e.g. the formation of andalusite in slates and schists).

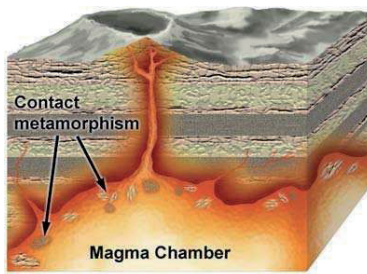


Fig. 69. Magma as a source of heat for Skarn deposits formation.

If carbonate rocks are present, skarn = tactite (a Ca-Mg silicate rock) is frequently formed by decarbonation and addition of silica. This process releases large quantities of CO₂ that may pass into the magma inducing profound changes. Massive ore bodies may occur in proximity to the skarn (proximal contact-metasomatic ore). The ore replaces carbonate rocks (or replaces the skarn) by a process called metasomatism (Fig. 70).

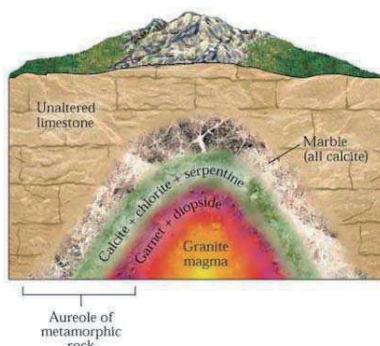


Fig. 70. Skarn deposits around carbonate.

The replacement is the result of the passage of hot aqueous fluids that are given-off by the cooling magmatic body or by dehydrating country rocks. If the metasomatic ore formation takes place at a distance from the intrusion, the ore will less likely be associated with skarn rock.

Skarn is an old Swedish mining term for a tough calc-silicate gangue that is associated with iron and sulphide ores. "Skarn" in USA commonly describes iron-rich rock bodies of Ca-Mg silicates formed from limestone or dolostone by abstraction of CO₂ and hydrothermal addition of SiO₂, Al, Fe and Mg in the contact aureole of intrusions.

• **How would ore bodies be formed in/around the Skarn?**

1. Emplacement of a hot magma body in cool country rocks causes the build-up of a thermal halo with outward migrating isotherms, driving-off water and other volatiles (Fig. 71).
2. During this prograde phase – contact metamorphism – the skarn area is born. In the skarn area, anhydrous minerals are formed that include grossular-andradite, diopside, forsterite and periclase (MgO, if dolomite was present), and part of the ore.
3. Outward from the intrusion, skarn is followed by a narrow zone of wollastonite and a shell of isochemical recrystallization of the precursor carbonate rock to carbonate marble.
4. The export of matter from the cooling magma into the country rocks is due to hot (maximum >700 °C) hypersaline melt, hydrothermal fluids and gas.
5. The hydrothermal fluid flow in aureole rocks has variable temperature and CO₂ concentrations. Commonly, initial heating will produce high CO₂.
6. Magmatic waters, i.e., hydrothermal fluids continue to exsolve from the intrusive magma during further cooling and deposit ores (Fig. 72).
7. The continuous hydrothermal fluids would transform the anhydrous mineral phases, such as MgO to hydrous phases brucite (Mg(OH)₂) and the formation of water-rich silicates (amphibole, epidote, talc, chlorite), concurrently with the main mass of the ore.

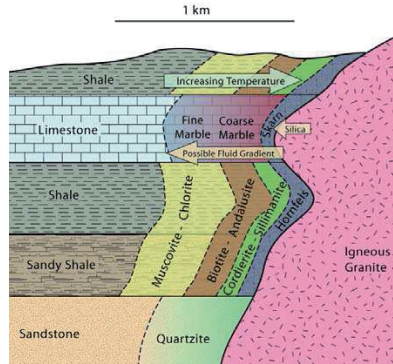


Fig. 71. Contact aureole around an igneous pluton.

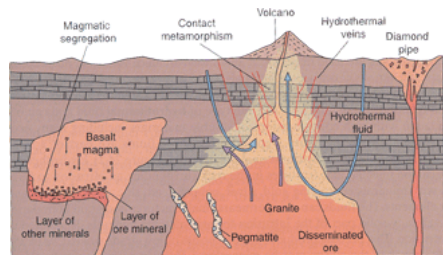


Fig. 72. Hydrothermal deposits form when water, heated by the cooling magma dissolves heavy metallic elements from the intrusion. These hydrothermal solutions cool as they leave. Slow moving solutions leave disseminated ore. If the solution cools quickly, it can deposit mineral rich veins.

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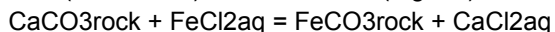
In this way Gold and Iron skarn deposits are formed from different hydrothermal solutions. Skarn orebodies display characteristically irregular outlines that can be explained by the two main factors, lithology and structures of the replaced host rocks, which impose chemical and physical controls on permeability and reactivity. Orebodies in skarns are often zoned, for example with copper in a proximal (near position) and lead-zinc in a more distal (far away) position. Skarn orebodies are a major source of many metals but also of industrial minerals including wollastonite, graphite, asbestos, magnesite, talc, boron and fluorite.

Other types of metasomatic ore deposits result from hydrothermal diagenetic and metamorphic fluids of evaporative and salt-solution brines. These hydrothermal fluids are Non-magmatic. Typically, the metasomatized rocks are marine limestones. This preference can be demonstrated with numerous examples (e.g. many lead-zinc orebodies (Fig. 73), gold as at Carlin, USA, magnesite and siderite).



Fig. 73. Lead-Zinc ores.

Main controls of the replacement process include the reactive surface and permeability of the precursor rock, pH and Eh of the mineralizing solutions, and the relative solubility of the participating minerals. The equation below describes the metasomatic formation of siderite rock (an iron ore) from limestone (Fig. 74).



In this case, cation exchange is the dominant mechanism, replacing each molecule of calcite with one of siderite.



Fig. 74. Siderite ore.

The emplacement of metasomatic ore is favoured by low-permeability rock horizons (e.g. shales) that form a physical barrier to upward flow (similar to petroleum traps). Focused hydrothermal solutions react more intensively with the carbonate host. Hydrothermal-metasomatic ore deposits are often

stratabound and occur in the same stratigraphic level across large regions (MVT deposits).

2.8 Porphyry Copper (Mo-Au-Sn-W) Deposits

Porphyry ore deposits are products of magmatic-hydrothermal activity at shallow crustal levels. Primarily copper, but also molybdenum, tin, tungsten and gold occur closely related to epizonal intrusions of porphyric magmatic rocks (Fig. 75).

Porphyries contain phenocrysts of hornblende, biotite, feldspar or quartz in aplitic (fine grained) groundmass. The phenocrysts crystallize when a fertile magma exists on deeper depths. Due to tectonic movements (subduction zones), the pressure is suddenly released and the magma is promoted to move to shallower depths and lose their volatile content. The fertile magma freezing during rapid ascent interprets the aplitic groundmass. Cu-Au-Mo porphyry ore deposits supply 75% of the world's Cu, 50% of Mo, nearly all of Re and 20% of gold.

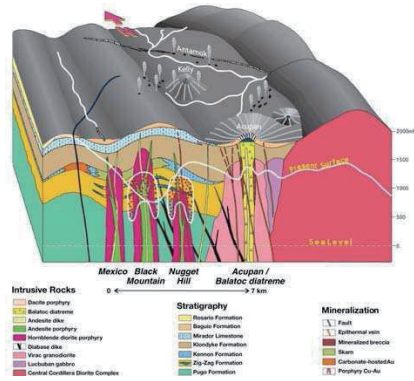


Fig. 75. Porphyry copper ore.

The most significant characteristics of copper porphyries include:

1. plug-like multiple porphyric intrusions below co-magmatic volcanoes, formed before mineralization;
2. an extraordinary tonnage of magmatic hydrothermal ore;
3. the ore occurs mainly in stockwork vein systems within the intrusion;
4. metal contents in ore are low to moderate, and supergene enrichment is often the key to exploitability;
5. extensive hydrothermal alteration and metals are vertically and cylindrically zoned in relation to the axis of the intrusion.

The major products from porphyry copper deposits are copper and molybdenum or copper and gold. The term porphyry copper refers to large, relatively low grade, intrusion-related deposits that can be mined using mass mining techniques (Fig. 76).

Host rocks of Cu-Au-Mo porphyry deposits are shallow (< 4000 m), sub-volcanic cylindric intrusions. The porphyry parent rocks are frequently:

1. calc-alkaline diorites (andesite-dacite),
2. monzonites (latite) or
3. granites (rhyolite) of I-type

Such parent rocks occur above subduction zones (convergent plate boundaries), either on active continental margins or in island arcs. Porphyries on active continental margins are marked by elevated Cu, Sn and Mo, those of island arcs are often Cu-Au. There are usually multiple episodes of intrusive activity, so, are commonly associated with swarms of dykes and intrusive breccias.



Fig. 76. Escondida Copper mine in Chile.

As stated before, the porphyry copper could be found in any country rocks, and often there are wide zones of closely fractured and altered rock surrounding the intrusions. The country rock alteration is distinctive and changes as mineralization is approached. Where sulphide mineralization occurs, surface weathering – by solutions - often produces rusty-stained bleached zones from which the metals – (Cu-Mo-Au) - have been leached; then re-deposited near the water table to form an enriched zone of secondary mineralization (Fig. 77).



Fig. 77. Porphyry Copper block.

• **Genesis of Porphyry Copper Deposits (PCDs)**

1. Various factors affect the porphyry copper formation, such as: magma type; volatile and oxygen content; the number, size, timing and depth of emplacement of mineralizing porphyry plutons; variations in country rock composition and fracturing; the rate of fluid mixing; density contrasts in the fluids; pressure and temperature gradients in the fluids, all combine to affect the porphyry copper genesis.

2. Fertile, volatile-enriched, highly oxidized magmas emplaced near highly permeable rock are the main sources of the ores. The ore could be crystallized/formed during the magma cooling and disseminated in the groundmass of the Porphyry intrusion (Figs. 78 and 79). This could happen due to the absence of reduced sulphur in the parent magma otherwise sulphide melts would form and lag behind the rising silicate liquid. However; porphyry copper ore deposits display a strong epithermal signature overprinting an earlier hydrothermal alteration. Therefore, two hydrothermal regimes are the most probable modes for ore formation

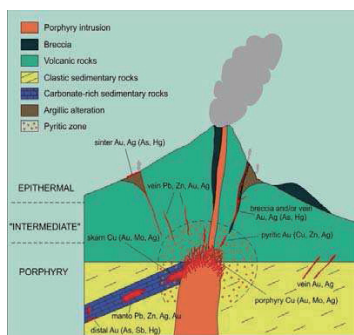


Fig. 78. Schematic diagram of a porphyry Cu system in the root zone of an andesitic stratovolcano. Figure shows mineral zonation and possible relationship to skarn, manto, "mesothermal" or "intermediate" precious-metal and base-metal vein and replacement, and epithermal precious-metal deposits.

3. Two hydrothermal regimes including i) orthomagmatic systems dominated by magmatic hydrothermal fluids (derived from molten rock) and ii) convective systems, dominated by non-magmatic meteoric waters (usually groundwater) interpret and portray the alteration and mineralization processes that have produced the wide variety of porphyry copper deposits.
- 
- The diagram illustrates a cross-section of a volcano. The top part is labeled 'Volcano' and shows a brown, conical shape with a crater at the top. Below the surface, the interior is shown in a lighter brown color, containing several small, dark, irregular shapes representing magma chambers or vents. A red line with arrows indicates the flow of fluid from the interior towards the base of the volcano. The base of the volcano is shown as a green area, representing the surrounding landscape or groundwater. The diagram is used to illustrate the two hydrothermal regimes mentioned in the text: orthomagmatic systems (dominated by magmatic hydrothermal fluids) and convective systems (dominated by non-magmatic meteoric waters).

4. **Orthomagmatic Model (regime/stage):** Volatiles and metals are concentrated during crystallisation of the magma, and then break through the crystallised carapace as metal-charged ore brines, i.e., magmatic hydrothermal fluids (Fig. 80). The initial wave of escaping hydrothermal fluids fractures the country

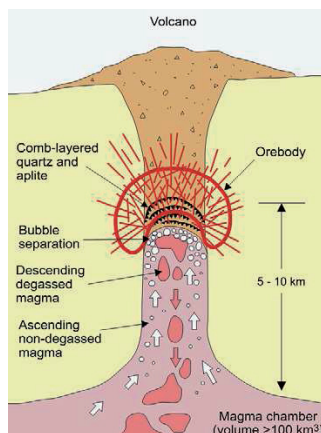


Fig. 79. Magma that is feeding a small sub-volcanic intrusion below a porphyry deposit.

rock and creates a crackle zone and plumbing system that controls the travel paths of subsequent hydrothermal fluids

and localizes alteration and mineralisation. Further cracking results from magmatic pressures, boiling and hydrofracturing.

5. Convective Model

(regime/stage): The convecting fluids (meteoric/groundwater-non-magmatic) transfer metals and other elements, and heat from the magma into the country rock and re-distribute elements in the convective system (Fig. 81). The two models are continuum.

The fundamental difference between them is the source and flowpath of the hydrothermal fluids. Here the hydrothermal fluid is originated from meteoric or seawater, i.e., non-magmatic. The permeability of the country rock is increased by the intrusive events to allow convective circulation. The convective hydrothermal fluids concentrate ore and gangue minerals near the intrusion.

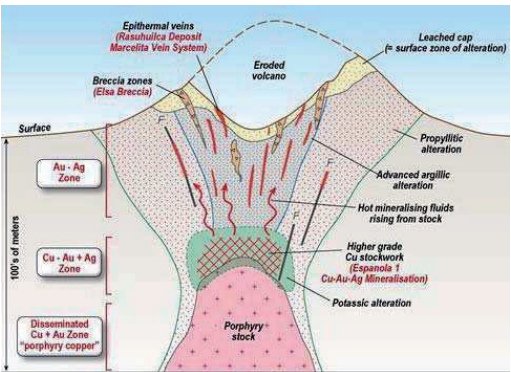


Fig. 80. Hydrothermal fluids concentrating the Porphyry Cu.

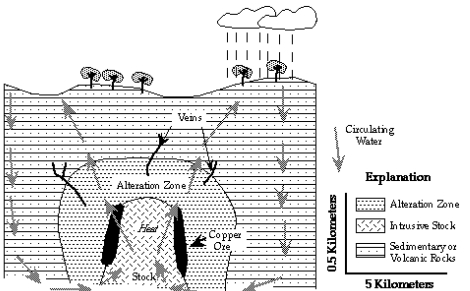


Fig. 81. Meteoric/groundwater convecting currents.

Current understanding (Fig. 82): Porphyry copper ore deposits mostly form beneath volcanoes, where magmas derived from a mid- to upper-crustal magma reservoir intrude into the shallower crust. Hydrothermal fluids transporting copper and sulfur (as well as other metals) are released from the magma and migrate into the surrounding and shallower crust. As the fluids cool and react with the surrounding rocks, metal-sulfide minerals are deposited and become concentrated in the crust. Recent research suggests

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that ore bodies may form during specific fluid-release events and by a combination of different processes (1–5) (Fig. 82).

Table 4. Orthomagmatic and convective models of porphyry Cu formation.

Orthomagmatic	Convective
Magmatic intrusion generates an ascending hydrothermal plume. Magmatic component constitutes up to 95% of the hydrothermal fluid	Permeable country rocks are the primary source of fluids. Magmatic fluids may be only 5% of the hydrothermal fluids.
Salinity is high, ranging from 15 wt % to 60 wt %.	Salinity is low, generally less than 15 wt %.
Multiple episodes of boiling, caused by repeated self-sealing and refracturing of the rocks.	Boiling is localised and of limited duration.
Fluid temperatures 400°C - 650°C, persisting over long periods of time.	Fluid temperatures may briefly reach 450°C, but quickly drop to around 250°C. The lower temperatures are then maintained for a long time.
Pervasive alteration and mineralisation form a series of shells around the core of the intrusion.	Alteration and mineralisation are both pervasive and fracture controlled.
Metals and sulphur are derived from the magma and are concentrated in residual fluids.	Metals and sulphur are scavenged from the enclosing rocks by convective ground waters.

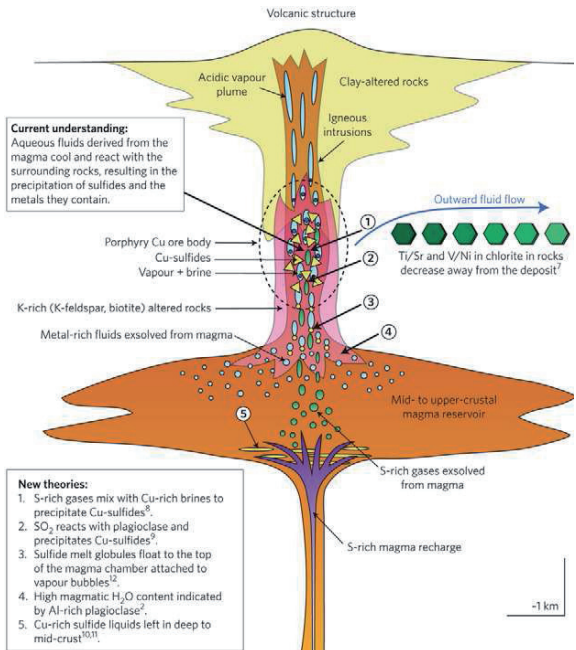


Fig. 82. Genesis of porphyry Cu.

• Mineralization of Porphyry Copper Deposits

Original sulphide minerals in porphyry copper deposits are pyrite, chalcopyrite, bornite and molybdenite. Gold is often in native found as tiny blobs along borders of sulphide crystals. Most of the sulphides occur in veins or fractures; most are intergrown with quartz or sericite. In many cases, the porphyry deposits have a central very low grade zone enclosed by 'shells' dominated by bornite, then chalcopyrite, and finally pyrite, which may be up to 15% of the rock. Molybdenite distribution is variable (Fig. 83). Radial fracture zones outside the pyrite halo may contain lead-zinc veins with gold and silver values.

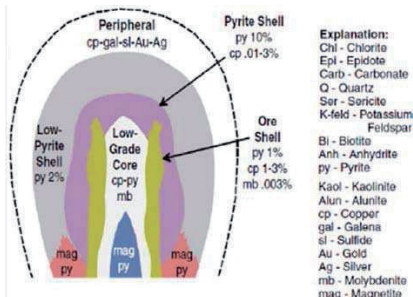


Fig. 83. Schematic cross section of ores associated with each alteration zone.

• Alteration zones related to Porphyry Copper Deposits

Alteration zones develop in/around granitic rocks with related porphyry deposits. If the alkali to hydrogen ratio is low, (i.e.; high K from the high volatile magma) feldspars, micas and other silicates become unstable and hydrolysis occurs when affected by hydrothermal solutions.

Four alteration types are common (Fig. 84):

1. Propylitic (weak hydrolysis): Quartz and alkali feldspar are stable, but plagioclase and mafic minerals react with the fluid to form albitised (Na) plagioclase, chlorite, epidote, carbonate and montmorillonite.
2. Argillic (more intense hydrolysis): Characterized by quartz, kaolinite and chlorite.
3. Phyllic: Quartz and sericite (fine muscovite), commonly accompanied by pyrite.
4. Potassic: High temperature alteration by concentrated hydrothermal fluids. All rock constituents are unstable. Alteration

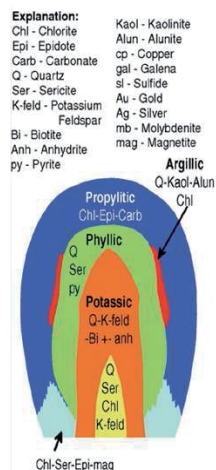


Fig. 84. Schematic cross section of hydrothermal alteration mineral zones, which consist of propylitic, phyllic, argillic, and potassic alteration zones.

assemblages of quartz (commonly resorbed), K-feldspar, biotite and intermediate plagioclase.

As a generalised model, these alteration assemblages form distinct zones around the intrusion, with a shell of potassic alteration grading outward through a shell of cream or green quartz and sericite (phyllitic), white, chalky clay (argillic) and then greenish chlorite, epidote, sodic plagioclase and carbonate (propylitic) alteration zones into unaltered country rock. In reality, the complete sequence is rarely developed or preserved, and assemblages are strongly influenced by the composition of the host rocks. Often there is early development of a wide area of secondary biotite that gives the rock a distinctive brownish colour.

- **Structural Features of Porphyry Copper Deposits**

Mineralization in porphyry deposits is mostly on fractures, so ground preparation or development of a 'plumbing system' is vitally important and

grades are best where the rocks are closely fractured. Porphyry-type mineral deposits (Fig. 85) result when large amounts of hot water that carry small amounts of metals pass through permeable rocks and deposit the metals. Intrusions associated with, porphyry copper deposits are porphyritic, reflecting rapid chilling. Porphyry dykes are very common and many breccia bodies reflect an explosive escape of volatiles (Fig. 86). Several periods of brecciation occur.

- **Classification of Porphyry Copper Deposits**

Porphyry copper deposits comprise three broad types (Fig. 85):

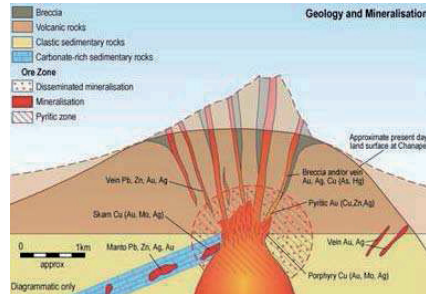


Fig. 85. Development of a 'plumbing system' within the porphyry copper plutons.

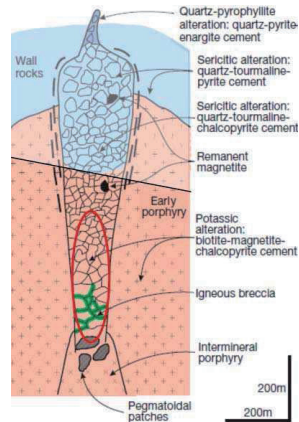


Fig. 86. Brecciation in porphyry copper pluton.

1. Plutonic: porphyry copper deposits occur in batholithic settings with mineralisation principally occurring in one or more phases of plutonic host rock.
2. Volcanic: occur in the roots of volcanoes, with mineralisation both in the volcanic rocks and in associated co-magmatic plutons.
3. Classic: occur with high-level, post-orogenic stocks that intrude unrelated host rocks; mineralization may occur within the stock, the country rock, or in both.

2.9 Hydrothermal vein deposits

Ore veins were the most important deposit type. More recently, the economic relevance of vein mining decreased compared to large-tonnage low grade operations such as those based on copper porphyries. Veins are tabular bodies of hydrothermal precipitates that typically occupy fissures. Less often, veins originate by metasomatic replacement of rock (replacement

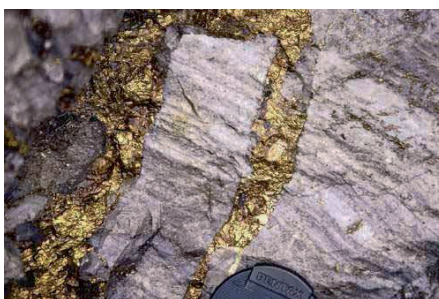


Fig. 87. Gold vein.

veins). Many veins develop upwards into a fan of thinner veins and veinlets which resemble a branching tree. Less than 0.5m thickness may allow profitable mining of high-grade gold and silver ore veins (Fig. 87), whereas tin and tungsten require a width of 1 m, barite and fluorite a minimum of 2 m. The world's longest veins may be those of the Mother Lode system of California, with 120km strike length. Most veins, however, have lengths between a few tens to several thousand metres.

Hydrothermal water which is a solvent and a transporter of dissolved metals deposit the minerals in the veins. The source of hydrothermal water is expelled from crystallizing magma, rain water that became ground water, seawater that intruded hot magmatic rocks near the mid-ocean ridges or could be water that was part of hydrous minerals that were metamorphosed to anhydrous phases which liberated the water into the pore space of rocks.

Mechanical properties of host rocks are the most important controls of vein formation, in contrast to metasomatic ore deposits that depend first on chemical properties. Fractures form more readily in competent rocks than in ductile material. Therefore, the cassiterite–quartz (muscovite, arsenopyrite, tourmaline) veins at Rutongo, Rwanda occur preferentially in vitreous quartzites and few cut across low-grade metamorphic schists.



Fig. 88. Highly fractured rock.

Very brittle rocks such as dolomite, rhyolite and quartzite are prone to form a network of short fractures instead of spatially separated longer ones. In that case, hydrothermal activity may result in stockwork ore (Fig. 89). “Stockwork ore bodies” consist of numerous short veins of three-dimensional orientation, which are so closely spaced (e.g. 10–30 veins/m in the tin deposit of Tongkeng-Changpo, South China) that the whole rock mass can be mined. Many vein deposits are associated with brecciated rock bodies that may host rich ore. As a rule, fissures and breccias have a much higher permeability than most consolidated rocks.

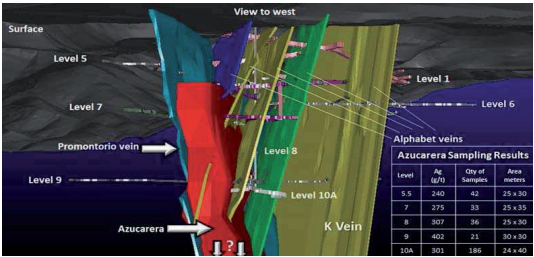


Fig. 89. Perspective view of a stockwork ore.

Many veins are associated with large-scale tensional tectonics including rifting (e.g. silver-lead-barite ore veins near the Rhine Rift) and late-orogenic relaxation of orogens (e.g. silver veins near Freiberg, Germany). Veins may also originate during convergent tectonics and folding thrusting (gold quartz veins in Western Australia) (Fig. 90).

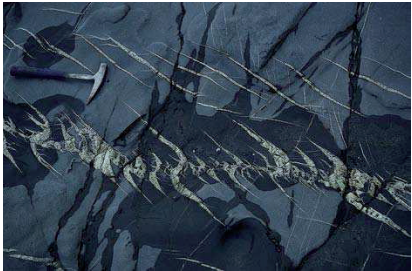


Fig. 90. Veins.

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During ore deposit formation, ore shoots (High-grade ore zones (Fig. 91)) were preferential channels of fluid flow and ore precipitation. Intersections of veins are often enriched, as are vein contacts with agents of metal precipitation in fissures, such as sulphides, organic matter and carbonates. The distribution of ore in veins is inhomogeneous and only a small part of the total vein fill may be exploitable. Related to either surface or volume, the ratio of economic to uneconomic parts of a vein is called the “coefficient of workability” (frequent values are 0.2–0.3).



Fig. 91. Ore shoot.

Hydrothermal ore veins of Cornwall are located in granites. The mineralization is not monophasic but the product of several activity periods. Tin, tungsten and much tourmaline are the main ore minerals. Highly saline aqueous fluid inclusions with tin, and cogenetic gaseous inclusions with CO₂ and W, Sn, suggest an origin by unmixing from high-temperature magmatic fluids.

2.10 Volcanogenic ore deposits (volcanogenic massive sulphides) (VMS)

The formation of a large number of important ore deposits is closely related to submarine and terrestrial volcanism (Fig. 92). The subvolcanic porphyry ore deposits and the mineralization related to the volcanic section of ophiolites (Cu-Zn-Au ore of Cyprus type) are types of volcanic ores. Here, other economically significant classes of volcanogenic ore deposits are introduced:

1. the submarine volcanogenic massive sulphide (VMS) deposits;
2. the terrestrial epithermal gold-silver-base metal deposits.

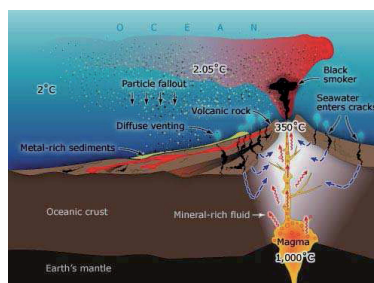


Fig. 92. Submarine exhalative” or “submarine hydrothermal”.

All volcanogenic deposits were formed near active volcanoes or subvolcanic centres, and occur in a “proximal” position.

Generally, ores that were formed by hydrothermal solutions venting on the ocean floor are called “submarine exhalative” or “submarine hydrothermal”.

“Submarine volcanic-exhalative deposits” (volcanic massive sulphides VMS are examples) are discerned from “Sedimentary-exhalative type” (Sedex). The first are clearly localized in volcanic centres, the second occur in sedimentation-dominated basins. In both cases, the rocks are called “exhalites” or “hydrothermal sediments”.

VMS

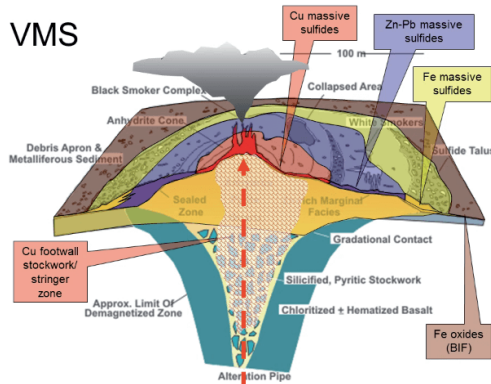


Fig. 93. Submarine exhalative” or “submarine hydrothermal.

2.10.1 Submarine massive sulphide (VMS) deposits

VMS deposits have been forming throughout geological history and they still are forming on the seafloor today (from black smokers). VMS are formed by submarine volcanic activity, or more precisely, sub-seafloor hydrothermal activity on the seafloor (Fig. 93). VMS-deposits with copper and zinc are hosted by sequences dominated by mafic volcanic rocks. VMS-deposits with zinc-lead-copper occur in sequences dominated by felsic volcanic rocks

sourced from continental crust and are best exemplified by the Kuroko deposits of the Miocene Green Tuff Belt of Japan.

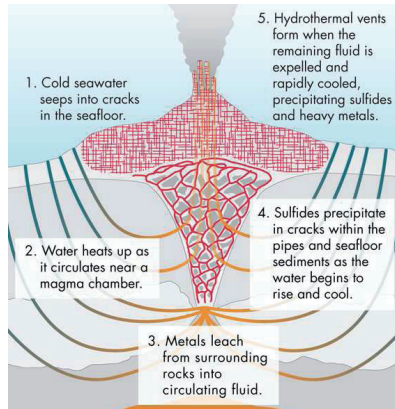


Fig. 94. Formation of VMS deposits.

• Genesis of VMS

VMS ore deposits are thought to result from seawater convection and hydrothermal dissolution of metals from pervaded volcanic rocks. It is believed that a link with the magmatic evolution and degassing of the volcanic rocks may complement the convection hypothesis (Fig. 94).

The VMS deposit form sulphide chimneys like those of the black smokers (old black smokers). Black plumes – hot waters - are emitted from volcanic vents/chimneys into the sea. VMS deposits are basically mushroom shaped tends to be copper rather than zinc rich. VMS deposits often form as clusters over a large intrusive heat source. If the heat chamber is long-lived, flat lenses of massive sulphide are formed. As the hydrothermal fluids reach the cold seawater, the temperature drops within seconds from 300 °C down to 100 °C and less. The fine clouds of sulphide cool and settles on the seafloor, building up a finely banded layers of pure sulphate (anhydrite) which are closest to the vent followed by copper, galena PbS and sphalerite ZnS. Beyond that the sulphur is exhausted and iron-sulphide/oxide and silica is all that's left to precipitate (Fig. 93).

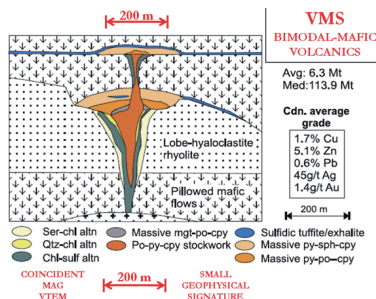


Fig. 95. Stratification of VMS.

The most important metals in VMS are Fe-Cu-Pb-Zn, with elevated trace contents of Cd-As-Sb-Bi, more rarely including gold and silver. Polymetallic ore deposits are commonly zoned with Fe-Cu at the base, followed upwards by Zn and Pb, and capped by barite, anhydrite or dense SiO₂ exhalites (chert, jasper) (Figs. 93 and 95). This chemical stratification can in some cases be explained by changes in the composition of the hydrothermal solutions. In other cases, a secondary mobilization of more easily soluble components (e.g. zinc) of early precipitates results in this pattern, caused by

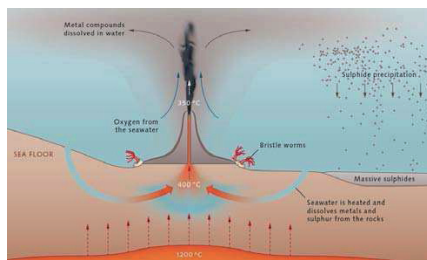


Fig. 96. Rapid cooling of the hydrothermal fluids.

continuous flow of hydrothermal solutions upward through the earlier metalliferous hydrothermal sediments ("zone refining").

Some of the metals are contributed by the underlying magma chamber, through the hydrothermal fluids (Fig. 96). It is the seawater circulation through the host volcanic that provides the remainder of the metal inputs. The secret of the high grade of the ore lies in rapid cooling of the hydrothermal fluid when it reaches the full seafloor.

• Forms and tectonic settings of VMS

VMS deposits show a great variety of forms, including lenses and blankets (the most common) (Fig. 93), but also mounds, pipes and stringer deposits. Colloidal textures prevailed initially, before zone refining, diagenesis and metamorphism caused coarsening and recrystallization. Banding, syndimentary soft-sediment deformation and graded bedding of sulphides are often observed in VMS deposits. Ooids and pisoids form where outpouring solutions caused both ore precipitation and agitation of hydrothermal seafloor sediment.

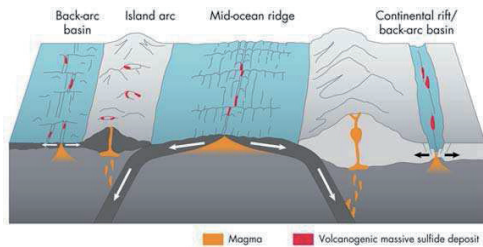


Fig. 97. Geologic settings where VMS are formed.

Iron-rich VMS occur predominantly with basalts, whereas Fe-Cu-Zn appear in volcanic terranes of andesite-dacites and Fe-Pb-Zn with rhyolites. The scarcity of Ti, V, Cr, Co and Ni in VMS deposits may be due to early precipitation of magnetite from mafic melt which extracts these metals. Generally, VMS occur in convergent plate boundary settings. However, the prevailing nature of the volcanic rocks and geochemical indices

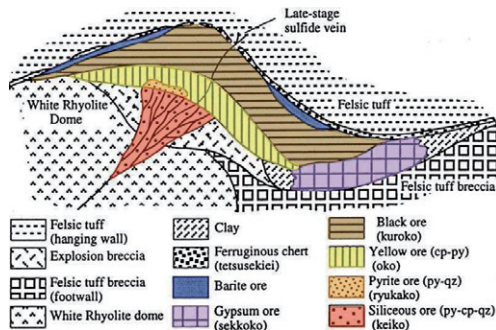


Fig. 98. Kuroko type VMS deposits.

imply that most VMS were generated during phases of major crustal extension, resulting in rifting, subsidence and deep marine conditions (Fig. 97).

- **Kuroko VMS deposits**

“Kuroko” ore deposits are an economically significant. The term “kuroko” is derived from the black lead zinc ore that was exploited in Japan for centuries. Many mines had also stockwork orebodies of yellow ore (“oko”) consisting of pyrite and chalcopyrite (gold). Fine-grained ores of ZnS, PbS, copper, pyrite and barite were formed during extrusion of rhyolite domes by hydrothermal exhalation through submarine vents and mound-building on the seafloor (Fig. 98). The temperature of the hydrothermal fluids rose during ore formation from 150 to 350 °C. Stable water isotopes indicate a seawater origin of the solutions with a low salinity of <5% NaCl (equivalent). Seawater contributed sulphur, but part of sulphide sulphur was leached from the magmatic rocks. The hydrothermal alteration in the footwall of the massive sulphide ores is concentrically zoned around the stockwork, from innermost K-feldspar and illite to distal montmorillonite and zeolites.

2.10.2 Terrestrial epithermal gold, silver and base metal ore deposits

Many important ore deposits are connected with terrestrial volcanism. In previous sections, iron oxide lavas and tuffs, and the subvolcanic porphyry copper ore deposits were mentioned. Here, we focus on a deposit class that is commonly called “epithermal”. These deposits formed by near surface hydrothermal processes (they occur at shallow



Fig. 99. Hot springs.

epizonal depth “100-300°C”; “hot springs” (Fig. 99)). Epithermal ore deposits are remarkable as a major source of gold, silver, zinc, lead, bismuth and the minerals alunite, barite and fluorite. Active epithermal systems of today’s volcanic regions are mainly investigated in view of their role as an energy source.

Epithermal ore bodies take the shape of veins, breccias, metasomatic masses and impregnations within hydrothermally altered host rocks. They are closely related to metalliferous sinter mounds that occur on the land surface ("hot springs deposits") (Fig. 99). Epithermal ore is formed in the aureole of near-surface subvolcanic magma bodies that induce a high heat flow and a steep geothermal gradient.

• High sulphidation epithermal deposits

Sulphur is an important component of magmatic volatiles. Near the surface, oxidation of sulphur produces strong acids that react with host rocks (Fig. 100). The resulting hydrothermal alteration is very conspicuous because of bleaching and formation of alunite. Alunite can also be formed when oxidic magmatic exhalations that are enriched in HCl and SO₂ dissolve in groundwater and react with host rocks. This is the environment of the formation of epithermal gold deposits of the alunite, or high sulphidation type. High sulphidation (high fS₂ in oxidic conditions) is indicated by ore minerals such as enargite, tennantite and covellite, and by advanced argillic alteration gangue minerals including alunite, pyrophyllite, dickite and kaolinite. High sulphidation epithermal deposits are often in close spatial association with copper-gold porphyries (Fig. 101).

• Low sulphidation epithermal deposits

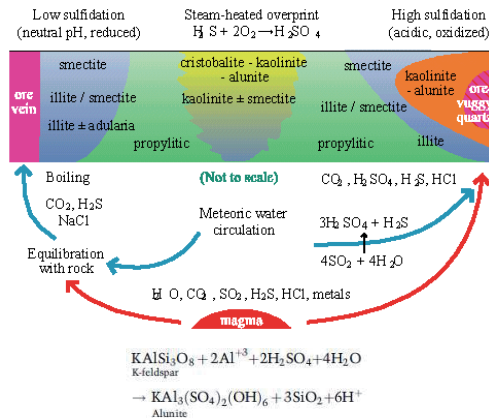


Fig. 100. Low and high sulphidation epithermal zones.

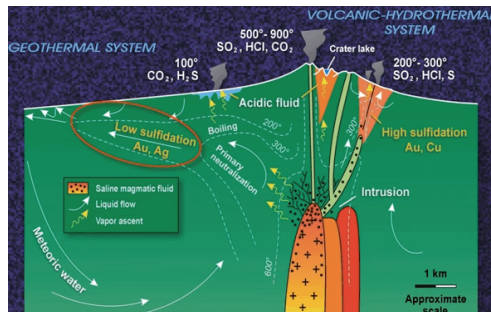


Fig. 101. Low and high sulphidation epithermal zones.

Chapter 3: Magmatic Ore Deposits

Epithermal deposits of the adularia-sericite, or low sulphidation type are produced by reduced and neutral to mildly alkaline fluids that carry H_2S and other reduced sulphur species (Figs. 100 and 101). In the low sulphidation type deposits, silver, gold and base metals dominate and typical gangue minerals are calcite, adularia and illite (Fig. 100). Generally, terrestrial volcanogenic deposits occur in convergent plate boundary and subduction settings. Many of these systems were triggered by tensional events.

Chapter 4: Supergene Ore Deposits

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1- Supergene Ore Deposits

Weathering of rocks and soil formation are the origin of many important ore deposits. Raw materials that are predominantly produced from supergene (secondary) mineral deposits include a diverse range of metals and minerals such as iron, manganese, aluminium, gallium, niobium and kaolin. Chemical weathering is dominated by reactions between minerals and rocks with meteoric water containing dissolved oxygen and carbon dioxide. The first reaction causes oxidation, for example Fe(II) to Fe(III), while the second reaction induces moderate acidity, especially after passage through a humic soil horizon that multiplies CO₂ concentration in seepage water (Fig. 1). Chemical weathering removes mobile cations (e.g. Ca²⁺, Na¹⁺, K¹⁺) relative to stable residual components (Al³⁺, Ti⁴⁺).



Fig. 1. Chemical weathering.

Site-specific conditions of pH and Eh are the main controls on reaction products. Weathering rates are a function of climate. Highest rates occurred when high CO₂ is available in the atmosphere with consequent warm and humid climate.

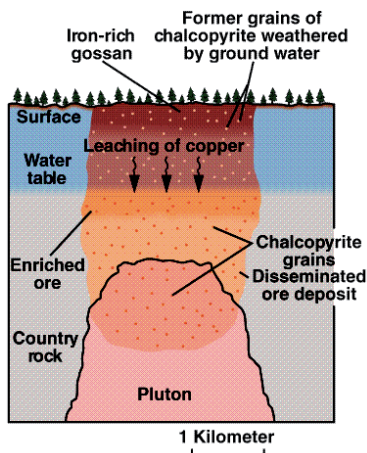


Fig. 2. Supergene enrichment.

The principle of supergene ore deposit formation is the concentration of some dilute but valuable component of the primary rock (Fig. 2). Two basically different process types may lead to concentration:

1. The valued component is enriched in a residuum, while much of the rock mass is dissolved and carried away.
2. The valued component is dissolved, transported and concentrated on re-precipitation.

In the latter case, the transport distance is commonly very short (metres to tens of metres). Some ore deposits, however, originate after long-distance transport dissolved in surface and groundwater. Examples include uranium in calcrete and in sandstone, iron in Tertiary river valleys of Western Australia and manganese in limestone karst caves.

An important special case of ore formation by weathering with metal transfer by meteoric seepage water is the supergene enrichment of un-exploitable low-grade primary mineralizations “protore”. Many copper, iron, manganese and silver deposits owe an economic ore grade to supergene enrichment processes.

Supergene ore deposits form in regions where weathering is favoured by a humid and hot climate

that promotes the profuse growth of vegetation (Fig. 3). Vegetation and

organic matter affect supergene alteration by two mechanisms. The first is the direct influence of plants on the soil water (e.g. retention). The second is the abundance of organic acids and of microbial activity that promote dissolution of primary minerals. In this environment, supergene, iron and aluminium have a very low solubility compared to alkalis and SiO_2 . Therefore, they are, iron and aluminium, enriched in the red, clayey-sandy soils of the tropics and subtropics that are generally called laterites. Ordinary laterites have little value except for making bricks and building roads.

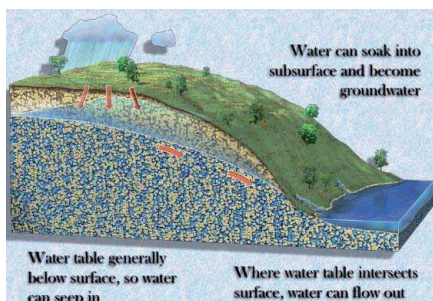


Fig. 3. Vegetation.



Fig. 4. Lateritic soil.

• Lateritic ore

Lateritic ore of (Al, Ni, Co, Fe, Mn, Cr, Au, etc.) is formed when exposure of suitable rocks coincides with favourable morphological, hydrogeological and

geochemical conditions that enhance dissolution, transport and precipitation. Time also has a role – long lasting weathering under stable conditions results in more mature and higher-grade ores. Laterite sensu stricto designates the upper leached and oxidized part of certain soil profiles with a well-developed vertical zonation (Fig. 4).

Lateritic regolith reaches more than 100m in thickness (Figs. 5 and 6). Principally, the lateritic regolith profile comprises an uppermost eluvial horizon (A), underlain by an illuvial zone (B) and the altered but still recognizable precursor rock in situ (C) that rests on unweathered fresh rock (R). Depending on the groundwater table, zone (B) is often divided into an upper kaolinitic and haematite mottled layer and a greenish-grey smectitic material below.

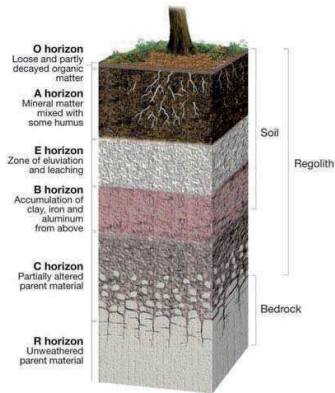


Fig. 5. Regolith structure.

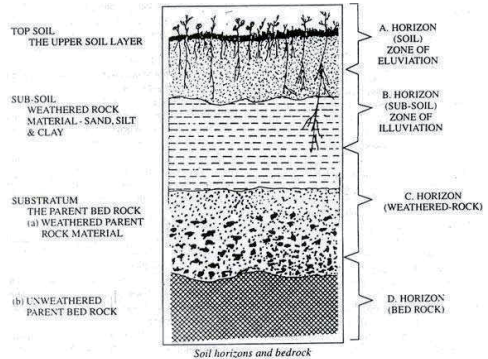


Fig. 6. Regolith structure.

In (A) or in (B), many laterites have hard crusts (duricrust) that are composed of SiO₂ (silcrete), iron (ferricrete) or manganese (mangcrete) oxides and oxy-hydroxides. Zone C can be crumbled by hand but often displays an upper clay horizon (feldspars are argillized) and a lower sandy horizon with intact feldspars but vermiculized biotite. The first is called “saprolite”, the second “saprock” (Fig. 5). Of course, bleaching indicates reduced conditions with dissolution and abstraction of iron, which is caused by organic acids derived from decaying vegetation.

- **Laterite zonation**

Laterite zonation is due to the effect of seasonal changes between dry and wet periods. During dry seasons, soil water moves towards the surface and evaporates, inducing alkalic conditions that promote silica solubility. In wet seasons, acidic rain water seeps downward, dissolves and re-precipitates matter or takes it out of the system. Textures of lateritic soils include banded crusts, shrinking fissures, concretions, pisolites and oolites, small vertical tubular structures, and many more. Newly formed solids are often amorphous and colloform, but diagenetic aging into minerals. An example is the transformation of iron hydroxide gel into goethite and haematite.

Textures of lateritic soils include banded crusts, shrinking fissures, concretions, pisolites and oolites, small vertical tubular structures, and many more. Newly formed solids are often amorphous and colloform, but diagenetic aging into minerals. An example is the transformation of iron hydroxide gel into goethite and haematite.



The crucial factor for effective leaching is water, modulated by hydraulic properties of soil and bedrock that influence the contact time between water and minerals (Fig. 7).

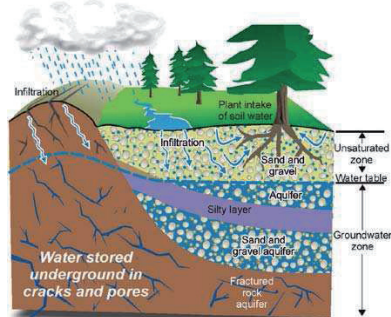


Fig. 7. Hydrogeologic system.

2- Residual (eluvial) ore deposits

In residual ore deposits, the economically interesting component is concentrated in situ, while weathering removes diluting parts of the rock (Fig. 8). Examples are residual and eluvial placers, bauxite, lateritic gold, platinum, iron (Ni, Co) and nickel ores, residual enrichment of sub-economic protore iron and manganese, and industrial minerals such as phosphate, magnesite and kaolin. The fundamental geochemical principle of the enrichment is the steady activity of a reaction front in soil (the valuable component is

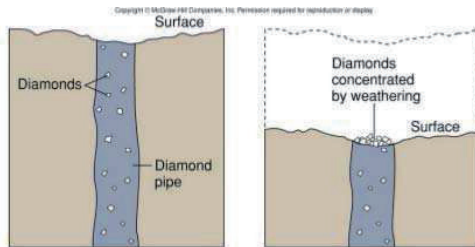


Fig. 8. land surface is lowered by weathering and erosion.

The fundamental geochemical principle of the enrichment is the steady activity of a reaction front in soil (the valuable component is

immobilized), while the land surface is lowered by weathering and erosion (Fig. 8).

2.1 Residual placers

Residual placers are concentrations of weathering resistant ore minerals that are due to supergene abstraction of non-ore material (gangue) (Fig. 9). The term eluvial (outwash) describes the same result, i.e., residual, although with more emphasis on physical removal of barren gangue. Many ore deposits, such as native gold, cassiterite and wolframite, contain heavy minerals that resist dissolution. In this case, physical weathering is the main agent. Residual placers are only stable in flat morphologies, because steeper slopes induce soil creep and with it a down-slope displacement of ore fragments (resulting in colluvial placers). Residual placers often occur at the “stone-line” level of a soil profile, where quartz and durable ore minerals are concentrated.

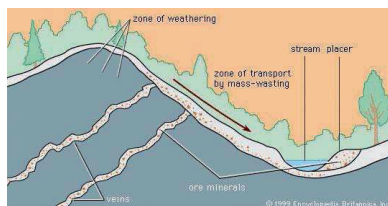


Fig. 9. Residual placer.

2.2 Bauxite ore deposits

Bauxite is the main source of Al in the world. Bauxite ore deposits originate either by in-situ weathering of aluminium silicate rocks (e.g. basalt; nepheline syenite), or by allochthonous sedimentation after erosion of bauxite soil blankets (Fig. 10). Autochthonous upland bauxites are discerned from lowland bauxites that include detrital (sedimentary) deposits. Residual bauxite horizons often extend over large areas and mark regional unconformities that are related to favourable tectonic and climatic conditions. Many aluminium silicate rocks contain an average of 15% Al_2O_3 that must be upgraded by weathering to at least 35% Al_2O_3 to be economically exploitable. Alumina enrichment is caused by leaching of SiO_2 and iron, two substances that are not easily dissolved.



Fig. 10. Bauxite ore deposit.

2.3 Lateritic gold deposits

Lateritic gold deposits are relatively recent discovery (Fig. 11). One of the largest representatives of this group was the Boddington bauxite mine in Western Australia, which until closure in 2001 was the biggest gold mine in Australia with an annual gold production of 2500kg. Pre-mining resources amounted to 60Mt of ore at 1.6ppm Au, apart from bauxite with gold contents of less than 1ppm.



Fig. 11. Laterite gold ore deposit.

Exploitable gold was located in near-surface, iron-alumina hard crusts that reached a thickness of 5m and in additional 8m thick lumpy Fe-Al laterite of the B horizon. Sources of the gold in soil at Boddington are quartz veins and hydrothermally altered bodies. Worldwide, numerous lateritic gold deposits are worked. They are attractive because exploration, extraction and processing of soil is less costly compared with hard rock mining.

2.4 Lateritic iron ore deposits

Lateritic iron ore deposits are not an important source of iron, because both deposit size and iron grades are rather low (Fig. 12). The ore consists of oolitic, red, yellow or brown haematite and goethite with elevated contents of H_2O , SiO_2 and Al_2O_3 . Most deposits take the form of autochthonous or locally transported hard crusts (ferricrete) that reach a thickness of only a few metres. Lateritic iron ore is derived from ultramafic source rocks and is exploited in several countries (Albania, Greece, Moa Bay, Cuba



Fig. 12. Laterite iron ore deposit.

and in the Philippines) as a high-iron limonite nickel ore (also termed “oxide nickel ore”) for pig-iron blast furnaces. In Fe-oxide nickel ore, nickel is absorbed in amorphous iron-hydroxides or occurs as inclusions in goethite. Residual supergene enrichment of iron pre-concentrations is the last upgrading event in the multistage evolution of high-grade haematite ore

deposits with 60–68 wt.% Fe, which could be derived from alteration of Precambrian BIF (Banded Iron Formation) by diagenetic brines.

2.5 Residual manganese

Residual manganese ore deposits are derived from rocks with above-average manganese contents. Laterites developed from such rocks include hard crusts (mangcrete) or earthy manganese ore that may blanket considerable areas (Fig. 13). In contrast to iron that tends to concentrate in the upper soil horizons (ferricrete), the slightly more mobile manganese is typically enriched in lower parts of the laterite profile.



Fig. 13. Laterite manganese ore deposit.

Lateritic manganese ores are often high-grade and of superior quality. Sources (protore) are manganese quartzites, carbonates and volcano-sedimentary rocks that contain exhalative-sedimentary pre enriched horizons.

3- Supergene enrichment by descending (vadose) solutions

3.1 Laterite Nickel deposits

Lateritic nickel ore deposits are of major economic importance. They form by intensive and long-lived tropical weathering of ordinary ultramafic rocks, which enriches nickel (and cobalt) (Fig. 14).

Two types of nickel laterite are distinguished:

- i. the oxide type in the upper, oxidized iron-rich part of the laterite; (Ni as hydroxide in the ferruginous zone) and
- ii. the silicate type in the lower, reduced saprolitic section of the regolith (Ni as hydrous silicates in saprolite).

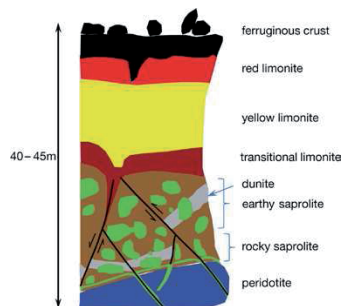


Fig. 14. Laterite Ni ore deposit.

Laterite regolith zonation and the redistribution of nickel are the result of meteoric water percolating through the soil. Olivine, pyroxene and serpentine are rapidly decomposed by acidic rain and soil water. From a surface value of 6, the pH of deeper soil water turns alkaline (pH 8.5).

Nickel is dissolved and flushed downward, where it is immobilized by ion exchange with Mg in newly forming magnesium hydrosilicates (Fig. 15). A schematic example of this reaction (neogenesis of the Ni-serpentine nepouite from serpentine) is given in equation.

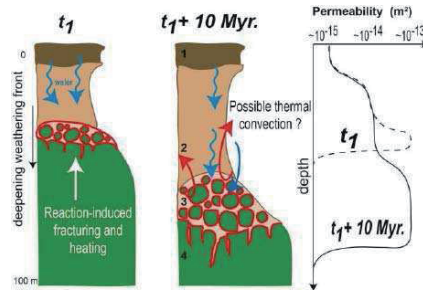
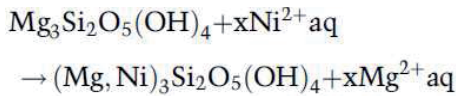


Fig. 15. Lateritization of the ophiolite.

Neogenesis of Ni-silicate in the saprolite zone:



Normally, groundwater removes the dissolved magnesium from the system. Some is precipitated as a carbonate (magnesite) in joints. New Caledonia is the largest ore province of Ni-laterites. The serpentinized and unaltered harzburgite-dunite displays an average 0.25% Ni (mainly in olivine) and 0.02% Co.

Lateritization of the exposed ophiolite resulted in a mature profile (Fig. 16):

1. Top: Massive goethitic (ferricrete); with low residual nickel concentration;
2. Limonite zone: with residual manganese, chromium, haematite and aluminium. Nickel, cobalt, magnesium, calcium and silica are leached and strongly depleted; in rare cases, residual nickel is exploitable;
3. Nontronite zone: Ferrallitic, earthy, red and yellow clays are the norm. Nickel and manganese are enriched to exploitable grades.
4. Saprolite zone: It is the main nickel ore horizon. Olivine and pyroxene are altered to colloform magnesium silicates that age into the minerals antigorite (serpentine), talc and smectitic clays; nickel is taken up by

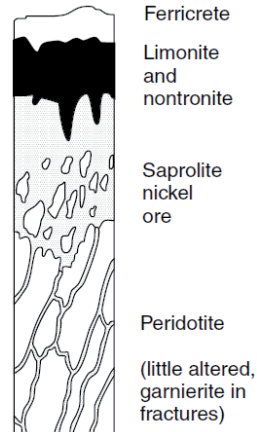


Fig. 16. Regolith profile of exploitable nickel laterite in New Caledonia ophiolite.

these minerals in their lattice by cation exchange of Mg^{2+} .

5. Bottom: Unaltered hard rock (protore) with rare garnierite in joints and fractures.

Supergene metal deposits form when common rock types or deeply buried primary ore bodies are exposed at or near the Earth's surface and undergo oxidation, dissolution and reconcentration of the metals. Supergene metal deposits are economically interesting because of their accessibility for extraction and increased grades. Scientifically they are attractive because of their mineralogical diversity and what they reveal about surficial history.

4- Supergene enrichment of pre-existing mineralization

Supergene, descendent alteration, may be exceptionally beneficial where pre-existing uneconomic mineralization or low-grade orebodies are affected (Fig. 17). Primary ore (protore) in deeper parts of the weathering profile is upgraded by seepage solutions. For sulphide copper and silver ores, iron oxides and certain uranium ore deposits, the supergene enrichment process is of economic significance. The supergene enrichment of sulphide ore is a consequence of near-surface oxidation of sulphides caused by meteoric water seeping downwards (Fig. 18). Other agents include dissolved oxygen and microbes that have a role as “self-replicating catalysts”. Sulphide-oxidizing microbes are cultivated in leach pads or large vats in order to decompose sulphide ores of, for example, gold, copper, nickel and cobalt.

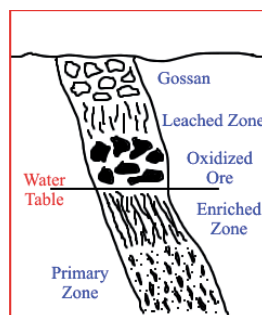


Fig. 17. Supergene enrichment.

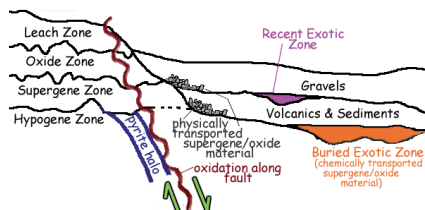
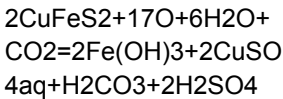
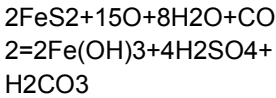


Fig. 18. Supergene enrichment.

4.1 Supergene enrichment of Copper Ore (Cu vein or Cu porphyry)

1. The oxidation of sulphur and iron in one mole of pyrite (FeS_2) (or marcasite) produces four moles of acidity. Consequently, weathering of pyrite bearing rock (and other copper bearing sulphides) is a natural or anthropogenic source of extremely acidic solutions (ARD, acid rock

drainage, or AMD, acid mine drainage) (Fig. 19). Supergene oxidation of pyrite-rich copper ore (pyrite and chalcopyrite) is illustrated by equations:



The results are acidic solutions, which move dissolved copper, but also silver, zinc and other metals through unsaturated soil down towards the groundwater (Figs. 19 and 20).

2. Because of their low solubility in oxic environments, iron hydroxides remain near the surface and form the characteristic botryoidal and cellular limonite-haematite masses of the "gossan" (Figs. 21 and 22). The gossan

can be called iron cap and is formed when weakly acid surface water percolates through the mineral deposit. Apart from iron (in addition to iron), manganese and gold are enriched in gossans. Traces of minor metals in the primary sulphides may be preserved, such as Ag, Pb, Cu, Zn, Cd, Mo, Ni, As and Sb. Gossans are often collapse breccias, because a large mass of the original ore rock was abstracted. Gossans are conspicuous indicators for the possible presence of hidden sulphide ore since the formation of limonite is accompanied by an increase in volume. Therefore, gossans are prime prospecting guides. Gossans can be orebodies in their own right, mainly of gold. Of course, not all supergene limonite-haematite masses are connected with valuable ore "false gossans". The distinction is possible by mineralogical and geochemical investigations.

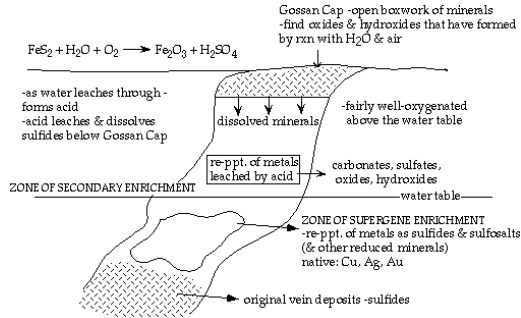


Fig. 19. Weathering of pyrite bearing minerals.

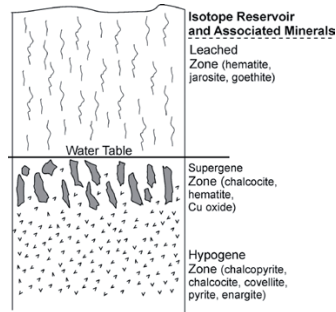


Fig. 20. Weathering of pyrite bearing minerals.

Chapter 4: Supergene Ore Deposits

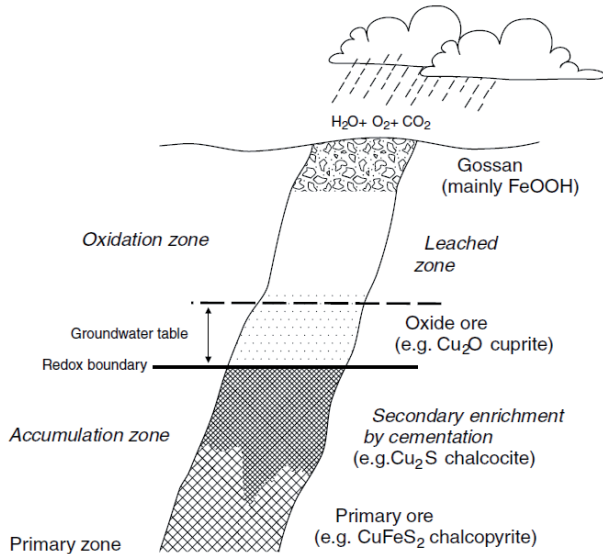


Fig. 21. Supergene enrichment of Copper ore from copper vein or porphyry copper pluton.

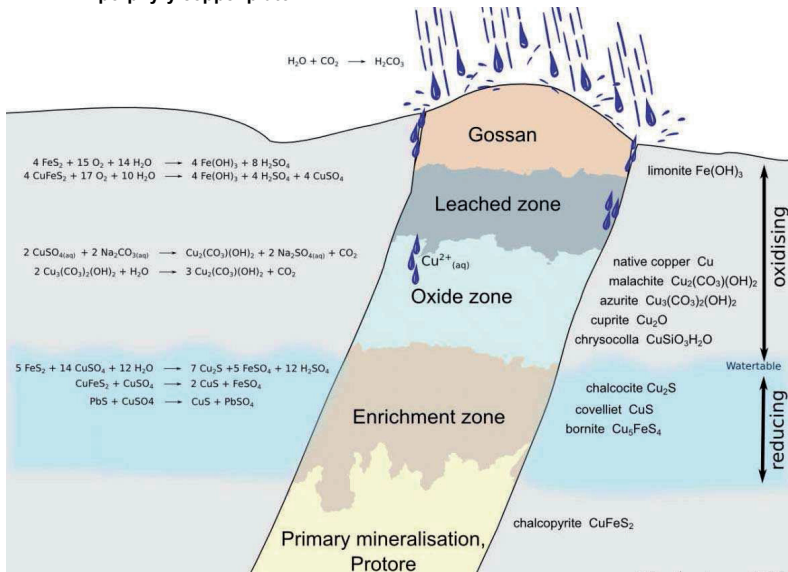


Fig. 22. Schematic view of a copper sulphide vein. The oxidation zone consists of the gossan, the leached zone and the oxidised zone. The reducing zone consists of the enrichment zone and the area of primary mineralization.

3. Below the gossan and above the groundwater table (leached zone/first oxidation zone), a zone of bleaching and silicate weathering (with kaolin, alunite, etc.) is often developed that reflects the former passage of strong acids (Figs. 21 and 22). The “leached” metal ions are transported down with the acidic solutions. Native copper and copper oxides (cuprite, tenorite), carbonates (malachite, azurite), sulphates (gypsum), chlorides (atacamite) and silicates (chrysocolla) could be precipitated in this zone and display a strikingly varicoloured paragenesis. This zone (leached zone/oxidation zone1) could be enriched to an exploitable volume and grade. Prominent examples are several Chilean copper porphyry deposits that have important “exotic oxide” outliers.
4. In the second oxidation zone (the groundwater area), where percolating acidified water ($\text{H}_2\text{CO}_3 + 2\text{H}_2\text{SO}_4$) meets the groundwater table, the active reaction front is reached. Here, extensive exothermic oxidation of primary sulphides (Chalcopyrite CuFeS_2) takes place, as well as the secondary enrichment by the aid of thermophilic microbes (Figs. 21 and 22). The secondary enrichment is referred to the possible oxidation of the CuSO_4 carried by the acidic solution and the deposition of CuO and Cu_2O as cementing oxides (Figs. 21 and 22). The efficiency of secondary enrichment depends mainly on conservation of the acidic environment (how much acidic water is found and for how long), but also on favourable hydraulic conditions (e.g. permeability). In the second oxidation zone as well, more noble metals (copper, silver, gold) could occur in the native form, which may be explained by a redox reaction: $2\text{Ag}^+ + 2\text{Fe}^{2+} = 2\text{Ag}^0 + 2\text{Fe}^{3+}$.
5. Below the water table, (actually below the Redox boundary), a secondary supergene enrichment zone (accumulation zone) is developed (Figs. 21 and 22). The greater part of the dissolved metal freight is re-precipitated in the reducing enrichment zone. This is achieved when the enriched CuSO_4 acidic solution is reduced – aided by the microbe actions - and the supergene sulfide phases - chalcocite, covellite and bornite - are deposited as cementing phases. Precipitation of enriched phases can be schematized as cation-exchange reactions:
 $\text{CuFeS}_2 + \text{CuSO}_4\text{aq} = 2\text{CuS (covellite)} + \text{FeSO}_4\text{aq}$.
 $5\text{FeS}_2 + 14\text{CuSO}_4\text{aq} + 12\text{H}_2\text{O} = 7\text{Cu}_2\text{S (chalcocite)} + 5\text{FeSO}_4\text{aq} + 12\text{H}_2\text{SO}_4$.

The ores of the enrichment zone may significantly surpass the metal content of the primary mineralisation.

5- Infiltration as an agent of ore formation

5.1 Uranium ore deposit

Infiltration is used to describe the movement of surface water into soil, porous rock, or karst. Infiltration ore deposits are formed when meteoric waters take up a substance that is dissolved by weathering, and concentrate it after considerable transport by infiltration in a different geological setting (Fig. 23). Uranium is a good example,

because during surficial alteration it is easily dissolved from granite, gneiss and felsic tuff, and is transported by creeks and rivers for kilometres, until infiltrating into an aquifer where reduced conditions cause precipitation and concentration. Selective weathering of different minerals may produce a pattern of spatial or temporal separation, for example by first leaching traces of uranium, copper and zinc from plagioclase and Fe-Mn oxyhydroxides, followed by barium, lead and SiO₂ when the more stable K-feldspar is decomposed. Copper, lead and silver ores embedded in haematitic sandstone suites “red bed deposits” may have originated in a way similar to infiltration deposits of uranium (e.g. the Transfiguration deposit in Quebec, Canada).

The concentration of solutes (solved material) in surface and groundwater is generally very low. Enrichment to ore-grade and an exploitable volume is only possible where a large mass flow is focused into a highly efficient filter (Fig. 24).

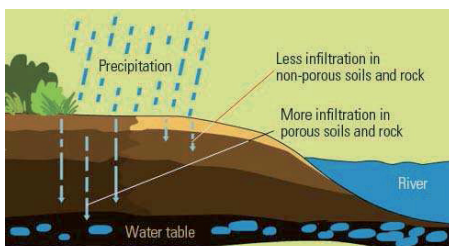


Fig. 23. Infiltration of ore bearing solutions in porous rocks.

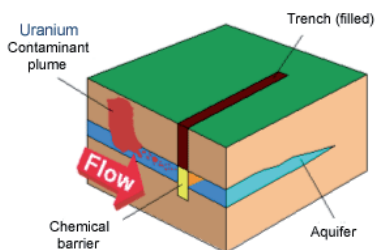


Fig. 24. Reactive barrier.

Geochemical barriers are most effective,

commonly in the form of a rapid change of pH and Eh. Geological actors include carbonates, H₂S or SO₄ in the pore waters, as well as organic matter. Because precipitation takes place in pre-existing rock, infiltration mineralization is clearly epigenetic process. Infiltration deposits of uranium are very common and economically important. Oxidative chemical weathering transforms uranium (IV) in rocks to uranium (VI), which forms complex ions with free SO₄, CO₃, OH, alkalis and humates. Under oxic conditions (low oxygen) in surface and groundwater, these complexes are stable and allow long distance transport. Geochemical barriers for uranium are phosphates, arsenates, vanadates and carbonates in percolated rocks (Fig. 25).

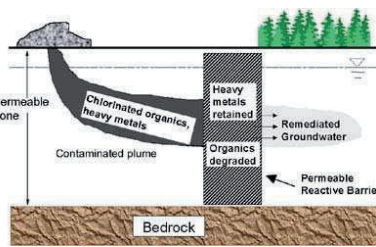


Fig. 25. Geochemical barrier.

Uranium infiltration mineralization and ore deposits occur in permeable sandstone and conglomerate (Colorado Plateau, or “sandstone type”), in volcanic ash beds, in faults and breccia bodies, in peat, lignite and coal seams, in asphalt, and in terrestrial calcite crusts of semiarid lands (the “calcrete type”). The genesis of the last includes a pronounced

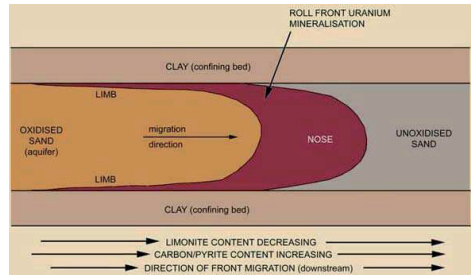


Fig. 26. Roll-front deposits cut across bedding. Uranium-bearing ground waters precipitate uranium oxide minerals when they come in contact with reducing conditions.

component of evaporation. The shape of orebodies varies widely, including tabular lenses, pockets and the characteristic “roll fronts”, often in connection with buried river courses (Fig. 26). Roll fronts mark the redox boundary in space and time where the infiltrating meteoric water lost its capacity to oxidize the percolated sandstone and to retain – the sandstone - uraniumin solution. Ore minerals include coffinite, uraninite, vanadium-rich clays and minor sulphides (of Fe, Mo, Cu, Pb, Zn, Se, etc.).

5.2 Karst systems

Formation of karst systems in carbonate rock bodies and the contemporaneous deposition of mineralization in the caves may also be related to infiltration processes (Fig. 27). Karst formation is caused by infiltration of meteoric water, which is enriched in CO_2 , for example by percolation through the organic soil horizon. Dissociation of carbonic acid produces acidity (H) that is essential in limestone dissolution. Because of a much slower reaction rate, dolomite is less affected by karstification

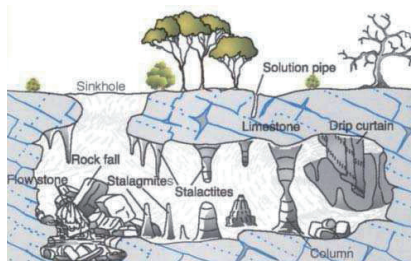


Fig. 27. Limestone Karstification.

There are three possible modes of ore deposit formation in karst:

1. Hydrothermal karst originating from hot ascending fluids. Structural relations such as cementation of host rock karst breccia by sulphides reveal simultaneous calcite dissolution and sulphide precipitation (Carlsbad caves in New Mexico (Fig. 28)) (note the possible connection with metasomatic ore formation).
2. Common meteoric karst, formed by downward percolating water with dissolved carbon dioxide, which was later filled by ascending hydrothermal fluids with ore and gangue, independent of karstification.
3. Common meteoric karst, and synchronous supergene infiltration of ore elements.

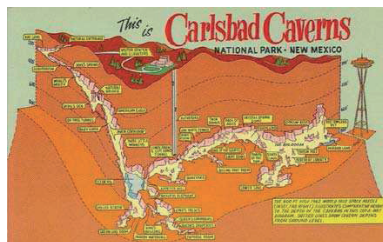


Fig. 28. Carlsbad caves.

Chapter 5: Sedimentary Ore Deposits

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1- Sedimentary Ore Deposits

The processes of sedimentation include physical, chemical and biological components. The prevailing regime lends its name to the major subgroups of sediments and sedimentary rocks: mechanical, chemical and biogenic sediments (Fig. 1). Sediments are also classified according to the provenance of the material into allochthonous and autochthonous. Allochthonous, are gravel, sand and certain clays. The source of the particles composing these rocks is distant from the deposits where the material was

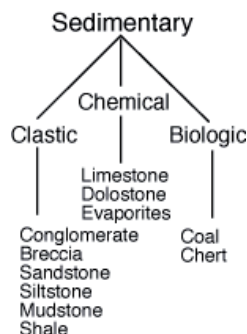


Fig. 1. Classification of sedimentary rocks.

transported. Metallic ore deposits of this genetic group are called placers. Placers are mechanical enrichments of heavy and chemically resistant native metals and minerals by flowing or otherwise agitated water.

Autochthonous raw material deposits were formed at the site of the deposit. This group includes chemical precipitates and partially biogenic substances such as:

- i. carbonates (limestone, dolomite, magnesite);
- ii. evaporites (rock salt, potassium salt, gypsum),
- iii. some massive and oolitic ironstones,
- iv. banded iron formations,
- v. marine phosphates,
- vi. sedimentary sulphide ores and manganese ore beds.

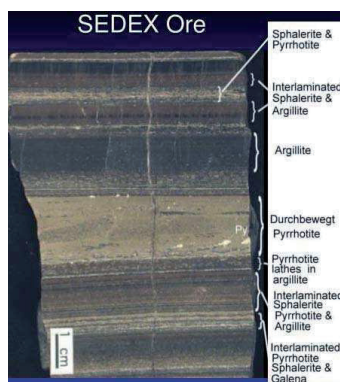


Fig. 2. SedEX.

Organic formation is the prevailing component for coal, oil shale, diatomite and for part of the limestones (e.g. chalk). In sedimentology, soils are also considered as autochthonous sediments. However, ore deposits related to soil formation processes were presented in the supergene enrichment. Many metalliferous sulphide, oxide, carbonate and sulphate deposits originate by autochthonous sedimentation following hydrothermal exhalation, precipitation of solutes and deposition of ore particles on the seafloor. Sensus

stricto, all ores with this origin are “sedimentary” (i.e., formed in a sedimentation-dominated environment and distal from contemporaneous volcanism. This is the class of sedimentary-exhalative (Sedex) deposits (Fig. 2).

Sedimentary ore formation will almost always have a biogenic component. In some cases this is very obvious, for example in phosphate deposits made of bones and coprolites, or a lignite seam composed of fallen trees.

2- Black shales in metallogenesis

Carbonaceous shales are important host rocks of sedimentary ore deposits (Fig. 3). They are fine grained laminated clastic sediments consisting variably of quartz, carbonate and clay with more than 1% of organic matter. The black shale sedimentary environment is characterized by low energy, benthic anoxia and often, by euxinic conditions (associated with stable reduced sulphur species in water above the seafloor).

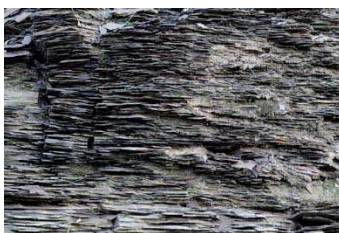


Fig. 3. Black shale.

The enrichment of metals in the black shales could be achieved by:

- I. the aid of organic matter which binds trace metals such as Ni, Co, Cu and Zn;
- II. reduction, such as V, U, Mo and Cr;
- III. high biological productivity, such as Ba, P and Cd.

Some black shale deposits are economically important such as the European copper shale and the Central African copper-cobalt shale. Other profitable use is the recovery of the energy inherent in organic matter, either by direct burning or by the production of synthetic oil.

3- Placer deposits

Sedimentary placer deposits are mechanically formed concentrations of heavy, durable minerals that may occur in



Fig. 4. Placer deposits.

Chapter 5: Sedimentary Ore Deposits

transported soil or regolith (colluvial), in fluvatile and in coastal sediments (Fig. 4).

Placer deposits of economic importance are classified as residual, eluvial, colluvial, fluvatile and coastal. Placers are important sources of gold (Fig. 5), platinum metals, tin, titanium (rutile, ilmenite), zircon, rare earth elements (monazite) and gemstones (diamond, garnet, ruby). Precondition of the concentration of placer minerals is their mechanical and chemical durability during weathering and transport (Fig. 6) and their elevated density compared with the ordinary rock forming minerals (Table 1).



Fig. 5. Placer deposits.

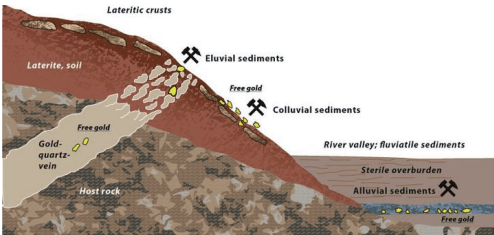


Fig. 6. Durability of placer deposits.

Table 1. Density of common placer and gangue minerals g/cm³.

Feldspar	2.5–2.8	Monazite	5.2–5.4
Quartz	2.65	Magnetite	5.2
Mafic silicates	ca. 3–4	Columbite	5.2–7.9
Diamond	3.5	Scheelite	5.9–6.1
Topaz	3.5–3.6	Cassiterite	6.8–7.1
Garnet	3.6–4.3	Wolframite	7.0–7.5
Corundum (ruby)	3.9–4.1	Cinnabarite	8.1
Rutile	4.2–4.3	Uraninite	7.5–9.7
Ilmenite	4.5–5.0	Platinum metals	12–21.5
Zircon	4.7	Gold	15–19.3

3.1 Colluvial placers

Colluvial placers originate by downslope creep of soil from weathered primary deposits (Fig. 7). Heavy minerals move to the base of the regolith, whereas lighter and fine grained fractions are displaced upwards. Ore

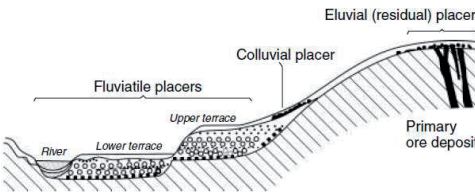


Fig. 7. Cross-section of a river valley near an outcropping primary ore deposit, which is the source of residual (eluvial), colluvial and fluvatile placers (black dots).

bodies are sheet- or channel-like bodies consisting of ore (e.g. cassiterite) and gangue (quartz) minerals, and of rock fragments. Colluvial ore minerals at the foot of a slope can be eroded by rivers and re-concentrated by alluvial processes.

3.2 Fluvatile/alluvial placers

Fluviatile/alluvial placers occur in active stream channels and in older river terraces (Fig. 8). Heavy mineral concentrations form at morphologically well-defined sites, mainly characterized by changes of flow velocity. Such sites include large boulders, rock bars, gravel beds on the inside bank of river bends, the downstream end of gravel and sand islands (point bars), but also reed patches. Hydraulic equivalence, *sensu stricto*, results in concurrent deposition of small high density grains with larger and lighter ones (e.g. fine-grained gold in coarse alluvial gravel). Geologically young alluvial placers are mainly exploited for gold.

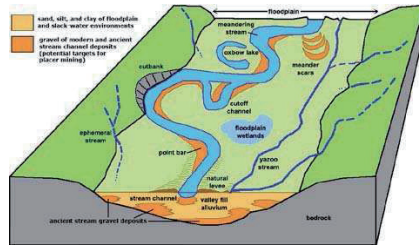


Fig. 8. Fluviatile/alluvial placer deposits.

3.3 Coastal placers

Coastal placers are mainly formed in surf zones (Fig. 9). In contrast to typical alluvial placers, both light and heavy minerals occur roughly in the same grain size that is almost exclusively sand. When a wave runs out, sand grains settle for a moment. Return flow to the sea is laminar and only light minerals can be entrained. Heavy minerals are enriched in a narrow linear strip along the beach. Other coastal processes may enhance placer formation including tides, lateral currents, wind and especially storms that induce higher waves.

Many promising coastal placer deposits of coarse-grained (90–300 mm) quartz and the heavy minerals rutile, zircon, ilmenite and



Fig. 9. Coastal placer deposits.

altered ilmenite (leucoxene) have been outlined. Coastal placers can consist of up to 80 wt.% heavy minerals. Their contribution of rutile and zircon to world markets is of high economic importance. They are lesser sources of diamond and cassiterite. Gold and platinum are rare in coastal placers.

4- Banded Iron Formations (BIFs)

Autochthonous iron ores are chemical, partly biogenic marine sediments. Banded iron formations (BIF) are layered, banded (0.5–3 cm) and laminated (<1mm) rocks containing >15% iron (Fig. 10). BIF consist of quartz and magnetite layers that form sedimentary units reaching lateral extensions of thousands of Km and a thickness of hundreds of meters.



Fig. 10. Banded Iron Formations (BIF).

BIF could be subdivided into three types:

1. Algoma type in submarine volcanic settings;
2. Superior type in marine shelf sediments; and
3. Rapitan type, which is closely related to glaciogenic marine sediments.

- **Algoma type**

Algoma ore BIF is found within volcanogenic sediments such as greywackes, tuffaceous and magmatic volcanic rocks (Fig. 11). Algoma ore beds are less than 50m thick and display a transition from oxide through carbonate and silicate to sulphide facies. Oxide facies iron occurs as magnetite, haematite is rare. In several districts, numerous exhalative centres with sulphides and some larger VHMS deposits punctuate Algoma BIF horizons. Algoma type BIF is formed on the seafloor by exhalative, hydrothermal-sedimentary processes driven by heat anomalies. Magmatic fluids may also participate in ore



Fig. 11. Algoma type BIF.

formation. Consequently, Algoma type BIF ore is essentially volcanogenic ore deposit.

- **Superior type BIF**

Superior type BIF are hosted in sedimentary sequences that include black, organic-rich and silicic shale, quartzite and dolomite, and transgress older basement (Fig. 12). Bimodal volcanic rocks (rhyolite and basalt) may occasionally form part of the country rocks. This indicates a sedimentary environment of stable continental shelves covered by relatively shallow seas with some extensional tectonic strain and associated volcanism. Superior type BIFs are marine sediments of global extension. They are preserved in remnants of marine basins that reach tens of thousands of square kilometres.



Fig. 12. Algoma type BIF.

Superior type BIFs contain iron in oxide, carbonate and silicate phases; sulphides are rare. Iron and SiO₂ were deposited as fine-grained ooze (or as a gel). Primary precipitates were probably amorphous SiO₂, ferric hydroxide Fe(OH)₃ and nontronite (iron(III) rich member of the smectite group of clay minerals) (near landmasses, oxic conditions), or precursors of magnetite, siderite and greenalite [(Fe)₂-3Si₂O₅(OH)₄] (far from land; in deeper, anoxic basins) (Origin of BIF comes later). Superior type BIF have an Fe₂O₃/SiO₂ ratio of 0.98 - 1.26 (which means high silica content), typically 25–45 wt.% Fe, <3% each of Al₂O₃, MgO and CaO, and small contents of Mn, Ti, P and S.

- **Rapitan type BIF**

Rapitan Type BIF were deposited immediately during glaciations (Fig. 13). During glacial events, land and oceans were widely covered with ice ("Snowball Earth" hypothesis) causing a nearly global anoxia and the presence of dissolved ferrous iron in seawater. Inter and post-glacial melting of the ice caps resulted in formation of glacial sediments, re-



Fig. 13. Rapitan type BIF.

oxidation of the oceans, and precipitation of ferric iron and manganese oxides. As marine sediments, Rapitan type iron formations are similar to Superior type BIF, but have little economic significance.

In all BIF-based iron ores, iron was enriched either by:

- i. supergene enrichment processes typically resulting in “martite-goethite ore” with 60–63 wt.% Fe; or
- ii. hypogene hydrothermal processes forming “high-grade haematite ore”, with 60–68 wt.% Fe.

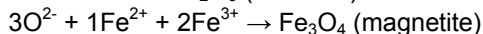
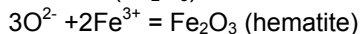
(Martite is a term that denotes haematite pseudomorphs replacing magnetite).

• **Genesis of Banded Iron Formations (BIFs)**

1. BIFs are sedimentary rocks dating back 3.8 to 1.8 billion years and have alternating layers of iron-rich minerals and a fine-grained silica rock called chert. About 3000 million years ago, there was no or very little oxygen dissolved in the oceans, because plants that produce oxygen had not yet evolved. However, the oceans contained a lot of dissolved silica, which came from the weathering of rocks.
2. This silica precipitated out from the seawater as layers of silica gel, which slowly hardened to become the rock chert (Fig. 14). Soluble iron oxide was also produced from:
 - i. submarine-exhalative systems similar to today's mid-ocean ridges;
 - ii. submarine-exhalative systems associated with intraplate tensional tectonic structures and volcanoes; and
 - iii. continental weathering.

Some 2500 million years ago oxygen-producing life forms started to evolve and oxygen became part of the Earth's atmosphere.

In time some oxygen also dissolved in the seawater where it reacted with the soluble iron oxide to form insoluble iron oxide. This precipitated out of solution on the ocean floor as the minerals magnetite (Fe_3O_4) and hematite (Fe_2O_3).



3. The repeated precipitation of silica and iron oxide - due to seasonal variation - resulted in the deposition of alternating layers of chert (which

is grey) (could be red jasper, a silica (SiO_2) with iron (Fe^{3+}) impurities), hematite (which is red) and magnetite (which is black) (Fig. 14). Thus, the name Banded Iron Formation BIF comes from the characteristic colour banding.

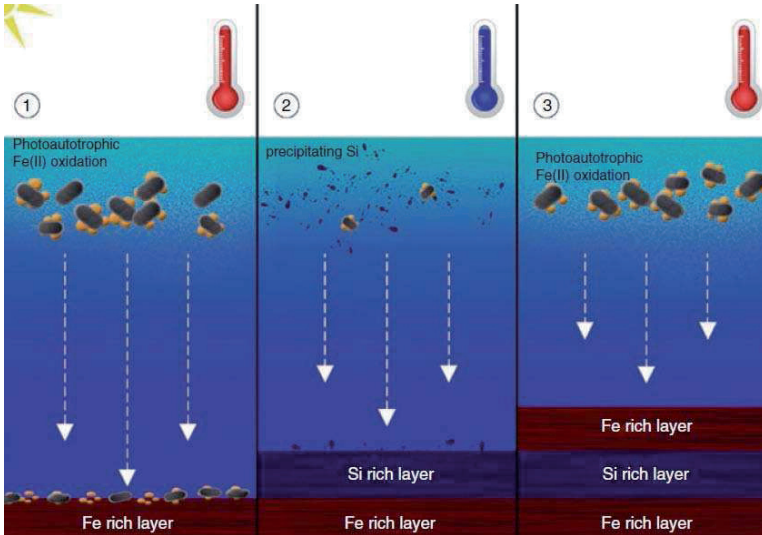


Fig. 14. Possible deposition of alternating iron and silicate mineral layers triggered by temperature fluctuations in the ocean water: (1) and (3) Moderate/higher temperatures yield relatively high photoautotrophic bacteria oxidation rates and thus iron(III) mineral formation. Therefore, biomass and Fe(III) settle together to the seafloor. (2) With decreasing temperatures photoautotrophic oxidation rates slow down and at the same time lower temperatures initiate abiotic Si precipitation from Si oversaturated ocean water. Si minerals then settle to the seafloor.

4. The source of oxygen is the stromatolites (Fig. 15). Stromatolites released free oxygen into seawater which immediately reacted with the iron which precipitated as iron oxide minerals. The stromatolites kept pumping oxygen, far outpacing the supply of iron (Fig. 15). Today, the ocean is full of oxygen, causing iron that reaches the ocean to instantly precipitate; therefore, iron deposits tend to be localized around black smokers (Fig. 15).

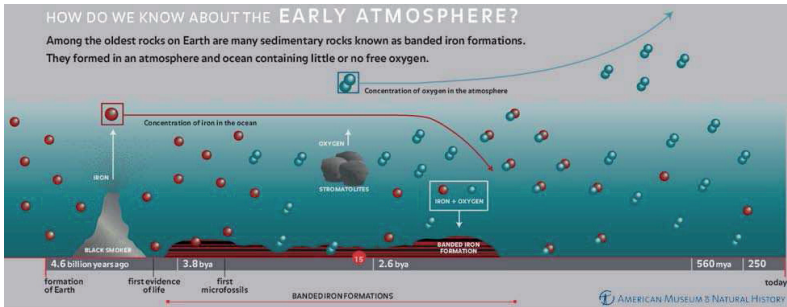


Fig. 15. Out pumping of oxygen by Stromatolites and the deposition of BIF.

5. At the Archean and Paleoproterozoic time, the earliest BIFs were also deposited in the absence of significant amounts of atmospheric oxygen (Figs. 16, 17). Anoxygenic phototrophic Fe(II)-oxidizing bacteria (photoferrotrophs and cyanobacteria) were the facilitators for the Fe(III) mineral deposition. The UV radiation and iron concentrations have made early marine environments inhospitable to life. The photosynthetic bacteria (photoferrotrophs and cyanobacteria) overcame these environmental conditions. High silica abundances in seawater was the initial survival of ancient photosynthetic bacteria, as well as to the early colonization of littoral marine environments, by forming Fe(II)/Fe(III)-Silicate complexes and nanoparticles in the ancient water column. These minerals/complexes acted as a 'sunscreen' against the high levels in incident UV radiation (Figs. 16, 17). By complexing with iron, silica would have lowered the level of soluble iron to more manageable levels, thus enabling the survival and evolution of early bacteria under high iron conditions.

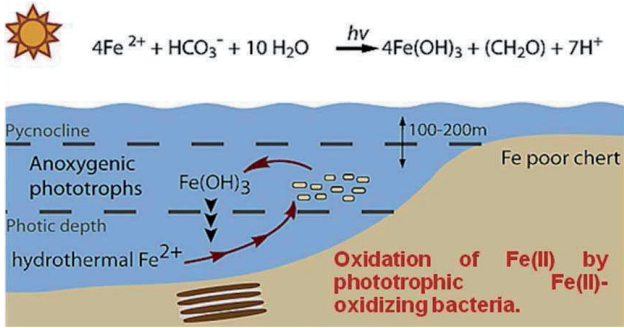


Fig. 16. Oxidation of Fe (II) by the aid of bacteria in the presence of sunlight for the deposition of BIF.

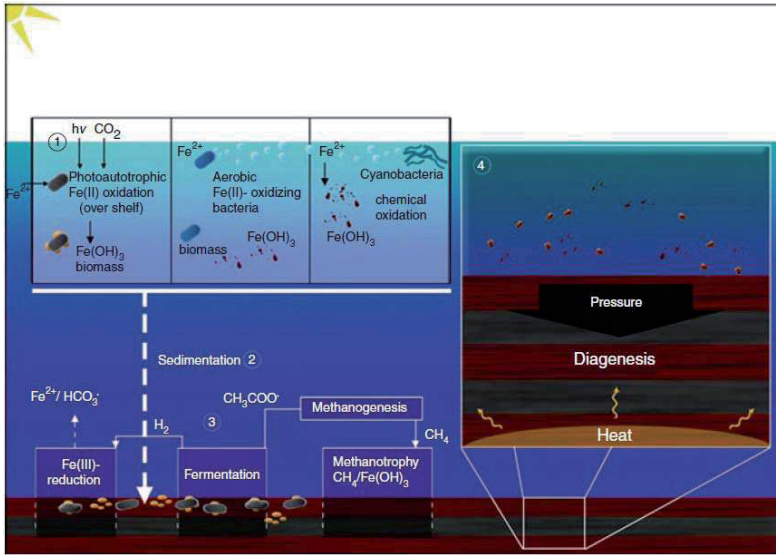


Fig. 17. Model summarizing potential biological and chemical processes during BIF deposition. (1) Hydrothermal Fe(II) is oxidized by photoautotrophic anoxygenic Fe(II) -oxidizing bacteria, aerobic Fe(II) -oxidizing bacteria or via chemical oxidation by cyanobacterially produced O_2 . (2) Biomass and Fe(III) sediment to ocean floor as cell-mineral aggregates. (3) After sedimentation metabolically driven redox processes by fermenters and Fe(III) reducers take place (and magnetite formation), possibly also involving methanogens and methanotrophs. (4) Pressure and temperature alter the source sediment and cause induced diagenetic/metamorphic overprint.

5- Oolitic Iron Ore

Formation of oolitic iron ore is restricted to Phanerozoic time. Economically outstanding mining led to the distinction of “Minette type” (limonitic) and “Clinton type” (haematitic) (Fig. 18). Always, these ores are sediments of shallow epicontinental seas (epicontinental sea is a shallow sea that covers central areas of continents during periods of high sea level that result in marine transgressions) or of shelf regions and are interbedded with clastic



Fig. 18. Oolitic iron ore.

sediments. Ore beds reach a thickness of about 30m and lateral extensions up to 150 km.

Oolitic iron ore is not banded and contains no primary colloidal silica. Textural varieties include mainly ooids and pisoids, but also fine grained iron ore particles in between (Fig. 19). Precipitation of ooids in agitated water is indicated by intraclasts in ooidal matrix, fractured and freshly mantled ooids, cores of ooids formed by quartz, phosphorite and fragments of fossils, oblique and cross-bedding. Iron oolite beds are often connected with rich fossil communities. Fossils are typically replaced and cemented by iron ore minerals. Single ooids consist of goethite, haematite, chamosite and siderite (with very little magnetite and pyrite). Iron and phosphorus are enriched $\text{Fe} = 34\%$ and $\text{P} = 0.6\%$.

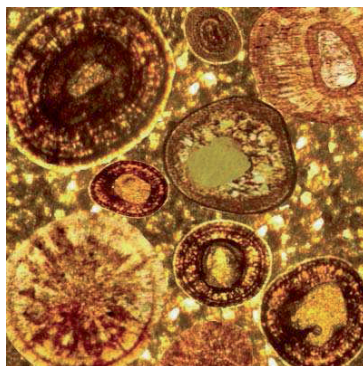


Fig. 19. Oolitic iron ore.

- **Genesis of Oolitic iron ore**

The coincidence of both sedimentary and diagenetic textures caused very different genetic interpretations. Originally aragonitic oolitic rock had been replaced by iron ore after lithification. However, many observations favour a synsedimentary supply of iron followed by chemical and biochemical precipitation. The unlithified sediments at the origin of oolitic iron ore consisted of ooids, pisoids and peloids that originated by rolling on the seafloor. The matrix between these larger grains consists of detrital minerals, fossils, organic material and fine grained iron-bearing kaolinitic clay. Subsequent diagenesis, mostly under reducing conditions, caused by bacterial decomposition of organic matter, typically leads to transformation of kaolinite into chamosite (an iron-rich chlorite) and neoformation of pyrite and siderite. Finally, many deposits were upgraded by oxidation caused by meteoric seepage waters. The source of iron in marine oolitic iron ores were probably lowlands experiencing intensive weathering in tropical climates, where humidity and profuse plant growth produced wetlands discharging river water rich in dissolved organic matter. Oxidized iron (Fe^{3+}) may be adsorbed to organic matter and clay (either in colloidal or particle size suspended in the

water). Mixing of river water with seawater (pH >7) provokes immediate precipitation of iron oxy hydroxides. Other possible iron sources that have been considered are coastal submarine groundwater springs.

6- Manganese Formations

Marine manganese oxide strata (manganese formations, MnF) (Fig. 20) occur commonly interlayered with Superior type BIF. The Kalahari manganese field comprises about half of the world's manganese resources. The concentration of Mn relative to Fe is believed to have been affected by earlier abstraction of iron from seawater and precipitation of BIF. Consequently, the soluble manganese was relatively enriched and only when concentration of O₂ continued to rise, Mn²⁺ was oxidized to Mn³⁺ and manganese rich beds were deposited. Manganese formations are only exploitable if enrichment processes acted on protore.



Fig. 20. MnF.

6.1 Oolitic manganese ore

Oolitic manganese ore occurs in seams within clastic marine sediments of epicontinental seas (Fig. 21). The metal source for oolitic manganese deposits may have been continental weathering or ocean-floor hydrothermal venting. Mn is dissolved in the restricted sea either abiotically or by microbial anaerobic methane oxidation:

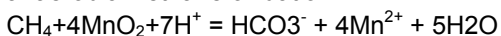
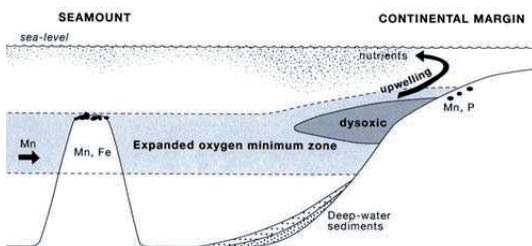


Fig. 21. Oolitic Mn ore.

This process leads to enrichment of anoxic deep water with dissolved manganese (Mn²⁺). Interaction of the Mn-enriched deep water with oxygen-rich marginal zones induces precipitation of Mn oxides, especially during transgressive phases (Black Sea is example).

More extensive reducing conditions on the seafloor could cause mobilization of Mn-oxides from deep-water sediments into the mid-water column, and their transfer to shallow-water



sediments where the upper boundary of the oxygen

Fig. 22. oceanic upwelling are associated with Mn-anomalies.

minimum zone intersects the seafloor (Fig. 22). At this position, Mn^{2+} can be re-oxidized to insoluble Mn-oxide, which falls to the bottom and could be converted to Mn-carbonate during early diagenetic reaction with organic matter. Thus, major periods of oceanic upwelling are associated with Mn-anomalies.

6.2 Manganese nodules

Manganese nodules occur over wide expanses of deep seafloor below the carbonate compensation depth (CCD). In shape and size, the Mn-nodules are compared to potatoes but, of course, smaller and larger specimens do occur (Fig. 23). Economically prospective manganese nodule fields include the



Fig. 23. Manganese nodules.

Clarion-Clipperton Zone in the Pacific. Nodules contain 29 wt.% Mn, 5% iron, 1.2% copper, 1.37% nickel, 1.2% cobalt and 15% SiO_2 . Mn-nodules have a botryoidal or smooth surface and consist of concentric shells of amorphous and crystalline manganese and iron oxyhydroxides.

Manganese nodules form on the deep seafloor as concretions formed by concentric layering of iron and manganese hydroxides around a core (Fig. 24). They grow at a very slow rate, on the order of millimeters per million years. The nodules formed by hydrogenetic processes are most abundant, these are formed by slow precipitation of metals from seawater. Mn-nodules and crusts have remarkable contents of valuable metals. The source of these

metals are most likely hydrothermal exhalations at mid-ocean ridges, submarine weathering of serpentinites, extraterrestrial dust and input from land to the oceans may contribute part of the endowment. Nodules are most abundant at depths between 4,000 m and 6,000 m in areas of low sedimentation rate. Hydrogenous nodules generally grow on siliceous and pelagic clay sediment. Manganese nodules grow when metal compounds dissolved in the water column (hydrogenous growth) or in water contained in the sediments (diagenetic growth) are deposited around a nucleus. Most nodules are a product of both diagenetic and hydrogenous growth.

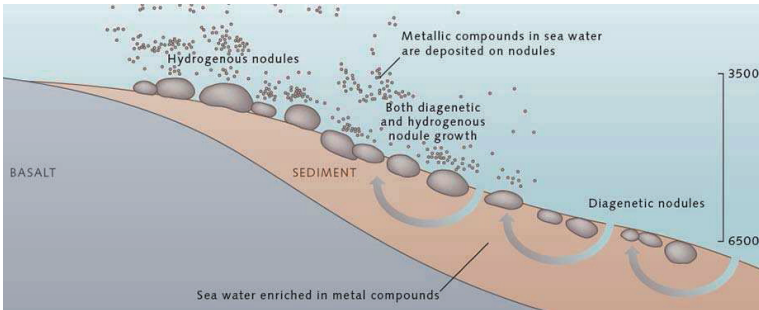


Fig. 24. Genesis of manganese nodules.

7- Sediment-hosted, submarine-exhalative (SedEx) base metal deposits

Hydrothermal-sedimentary sulphide ores occur in a continuum from:

- a proximal position to submarine volcanoes;
- more distal positions where only sparse ash layers in the sedimentary column point to synchronous volcanism (Fig. 25); and
- purely sedimentary settings.

Only the cases ii) and iii) are referred to as sedimentary-exhalative or short, SedEx type ore deposits. The first are volcanogenic massive sulfides (VMS).

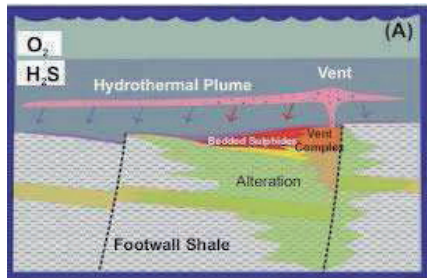


Fig. 25. SedEx from sparse ash layers in the sedimentary column.

The term SedEx implies syngenetic origin of ore and host rocks. The characteristic setting of these deposits is submarine rifting within

epicontinental seas. Ore bodies are typically related to faults, synsedimentary tectonic activity and the formation of local sub-basins. Extensional restricted basins associated with crustal thinning and distal volcanic centres may be part of the environment.

• Genesis of SedEX

1. SedEX deposits are generally formed in fault-bounded sedimentary basins on continental crust (within-plate locations) rather than in volcanic piles

on oceanic crust. To achieve SedEX deposits, the basin needs to accumulate several kilometres or tens of kilometres of oxygen lacking sediment, usually shales (Fig. 27). SedEX are close to or enclosed in organic rich sedimentary rocks (e.g. black shales) indicating a state of partial (dysoxia) or severe oxygen depletion (hypoxia to anoxia) of bottom waters (Fig. 28). The correlation between the black shale and the depletion of oxygen could be interpreted in terms of the outpouring of profuse hydrothermal metalliferous fluids causing an euxinic marine sub-basin environment (i.e. anoxic, highly reduced and H₂S stable).

2. The sulphides in SedEX were deposited in a geologically very short time. However, the clastic sedimentation rate in the host basins was certainly not exceptionally low, but the sulphide deposition rate was very high.
3. Fluids involved in SedEX ore formation include:
 - i. deeply convecting seawater and evaporitic brine (Fig.27);

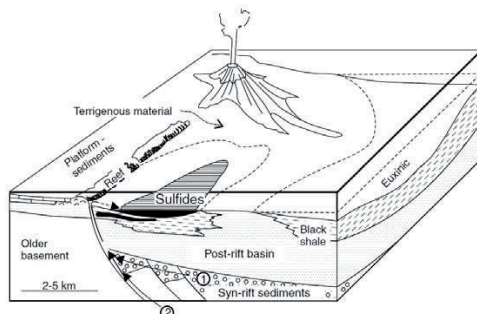


Fig. 26. Schematic sketch of the geological setting of an exhalative-sedimentary (SedEX) sulphide (barite) ore deposit in a black shale basin. Note distal rift-related basaltic and felsic volcanic centres.

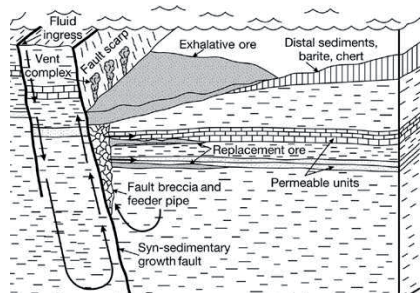


Fig. 27. Black shale containing SedEX.

Chapter 5: Sedimentary Ore Deposits

- ii. occluded formation fluids;
 - iii. diagenetic water of basinal sediments; and
 - iv. fluids mobilized from a deep source (e.g. elevated heatflow from the mantle through fault planes (Fig. 28)).
4. The heat derived from the hydrothermal solutions together

with the heat from the sediment burial leach the metals; lead, zinc and silver from the sediments and concentrate them with those in the hydrothermal solutions forming the SedEX (e.g., Broken Hill, Australia).

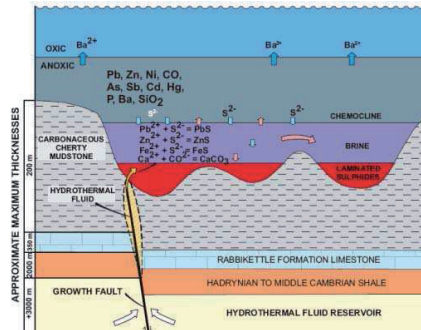


Fig. 28. Hydrothermal solutions as feeder for metals.

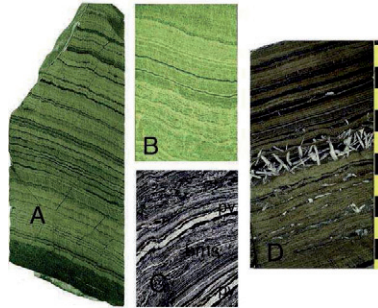


Fig. 29. SedEX.

Sedex ore bodies are stratiform and stratabound, and consist of massive, laminated or banded sulphides with varying admixture of clastic matter, barite and other exhalites (SiO₂, haematite, etc.) (Fig. 29). In appearance they are very similar to volcanic-hydrothermal exhalative (VMS) orebodies; the decisive difference is the sedimentary as opposed to the volcanic genetic setting. VMS are principally controlled by volcanic point heat sources, however, Sedex ore is commonly localized by extensional faults (e.g. basal growth faults) (Figs. 27 and 28).

Single ore beds are either monomineralic or mixtures of galena, sphalerite, pyrite and pyrrhotite, rarely

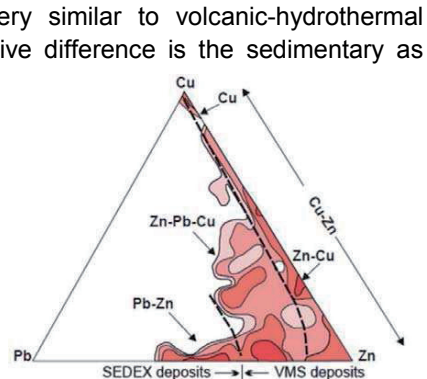


Fig. 30. SedEX and VMS.

Chapter 5: Sedimentary Ore Deposits

including more than traces of arsenopyrite and copper sulphides (Fig. 30). SedEx compete with porphyry deposits for the role of the largest base metal accumulations on Earth.

Chapter 6: Diagenetic Ore Deposits

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1- Diagenetic Ore Deposits

Diagenesis is “the sum of all chemical, physical and biologic changes undergone by sediment after its initial deposition, and during and after its lithification, exclusive of surficial alteration (weathering) and metamorphism (Fig. 1)”.

Examples of ore deposit types that illustrate the diagenetic realm include:

1. the European Copper Shale,
2. many deposits of the Central African Copper-Cobalt Belt,
3. Mississippi Valley Type (MVT) lead-zinc (barite-fluorite-celestite) deposits, and
4. orebodies that originate from brines derived by interaction with salt rocks (e.g. lead zinc and siderite in Northern Africa).

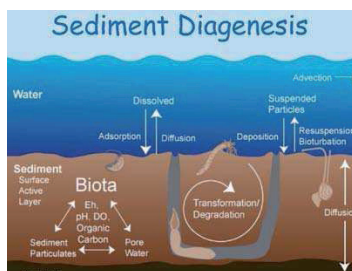


Fig. 1. Diagenetic processes.

The core of a diagenetic ore formation model is the argument that most sediment includes a large mass of water at the time of deposition “connate/formation water” (Fig. 2). During diagenesis, most of this initial water is expelled by mechanical compaction. When newly deposited, sand may contain 40 vol. % of water (connate/formation), clay 90%, and carbonates 50%. By diagenesis, the pore space occupied by water is reduced to < 1 vol.%. All this connate/formation water must leave the system by flowing upwards and to the margin of sedimentary basins.

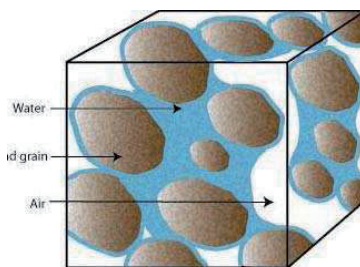


Fig. 2. Connate water.

• Characteristics of connate/formation water

1. Connate/formation water in sediments frequently attains temperatures of more than 100 °C and reaches a maximum of about 300 °C. By reacting with rocks, the connate/formation water acquires a high content of dissolved matter. When large masses of connate water

passee through geochemical or physical “traps”, minerals are precipitated (epigenetically) as ores.

2. Typical connate/formation waters are brines dominated by Na-Ca-Cl and elevated traces of HCO_3 , K, I and Br. (NB. brine sensu stricto is defined by a salt content of more than 10% (Table 1)).
3. Compared with seawater, iron, manganese, copper, zinc and many other elements are enriched in the connate/formation water. Magnesium and sulphate are depleted; the first by neogenesis of dolomite and chlorite (minerals containing Mg), the second by bacterial (Fig. 3) or thermochemical sulphate reduction to H_2S (Fig. 4).

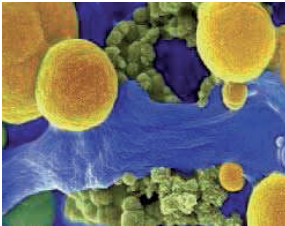


Fig. 3. Bacteria.

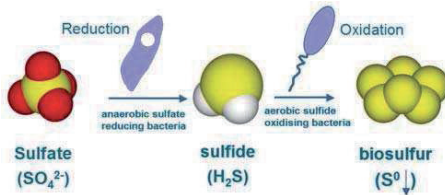


Fig. 4. Reduction of sulphate into sulfide.

4. Live microbes exist to a depth of several kilometres below the surface and to temperatures of $140\text{ }^{\circ}\text{C}$. In the presence of microbes, the organic matter is converted to CH_4 and CO_2 to produce the methane of biogenic gas deposits (methanogenesis) (Fig. 5). In the near-surface realm of early diagenesis, methanogenesis and sediment dewatering combine to cause methane rich cold seeps (cold enough

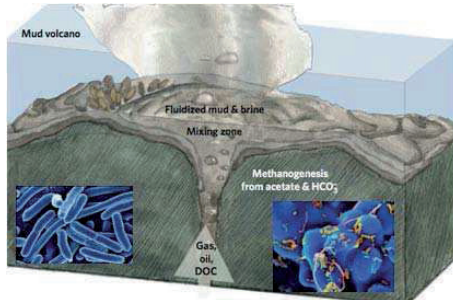


Fig. 5. Methanogenesis by the effect of bacteria.

formation/connate waters) that are sites of submarine barite deposit formation.

5. In the absence of microbes, at higher temperature ($>80^{\circ}\text{C}$) and greater depth, abiotic sulphate reduction based on destruction of organic matter (here represented by CH_4 and CH_2) is possible. By disintegrating anhydrite (SO_4), this process can cause a relative enrichment of CaCl_2 in the fluids.
6. NaCl may be retained and concentrated by clay beds (acting as a semi-permeable membrane). Probably more often, NaCl brines originate by evaporation at the surface or by dissolution of salt rocks in the basinal sediments. Mature, commonly deep connate/formation waters are mostly concentrated NaCl brines.
7. The oxygen of the connate/formation water is consumed by oxidation of Fe(II) . Therefore, most connate/formation waters display a low redox state, favouring high solubility of many metals. In solution, some metals may occur in the form of organo metallic complexes (Ni , V , part of Zn and Cu), while others such as Pb , Fe and most Cu and Zn are dissolved as simple ions.
8. Not all connate/formation waters are reduced, oxidized connate/formation waters are commonly enriched in SO_4 . This is possible when the connate water source rock includes predominantly haematitic sandstone, mudrock, limestone, anhydrite and halite. Oxidized fluids are able to dissolve Cu , Pb and Zn , to transport the metals over wide distances and to precipitate ore at suitable traps such as redox fronts established by organic-rich shale (e.g. the European Copper Shale). Barium and iron, in contrast, are insoluble in oxidized fluids.

- **How would connate/formation waters play their role in ore formation?**

Metal uptake (leaching by connate/formation water) in the source region (reaction zone=interface between connate/formation water and rock) or along connate/formation water flow paths is a function of kinetic and thermodynamic factors, including rock and water chemistry, T , P , Eh and pH .

In contact with formation waters, several common minerals are unstable, including amphibole, pyroxene, olivine and epidote. Their alteration releases trace metals into the aqueous phase.

Other mechanisms of liberating metals include ion exchange with clay minerals and simple diagenetic transformation of clays. An example is lead that is often adsorbed on kaolinite, but is released into solution (leached by solution) during/when illite replaces kaolinite.

The flow paths of expelled deep connate/formation water are determined by hydrogeological parameters such as permeable rock units, tectonic disaggregation and the general pressure gradient. Ore formation will only occur where a large mass of hot brine/connate water is focused into preferential flow paths, which may be envisaged as streams within permeable rocks or tectonic structures. In these channels, a rapid drop of temperature or pressure, reaction with host rocks, and mixing with chemically different water may precipitate solutes (ore minerals) in a small rock volume. Near-surface faults, reefs or karst cavities at the margins of sedimentary basins or basement islands

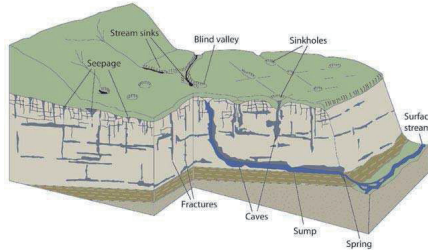


Fig. 6. Karstification.

often locate ore deposit formation (Fig. 6). In these channels, a rapid drop of temperature or pressure, reaction with host rocks, and mixing with chemically different water may precipitate solutes (ore minerals) in a small rock volume. Near-surface faults, reefs or karst cavities at the margins of sedimentary basins or basement islands

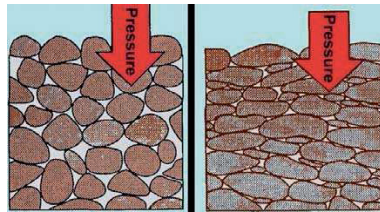


Fig. 7. Dewatering by compaction.

often locate ore deposit formation (Fig. 6).

The factors affecting the flow path of the diagenetic connate/formation waters are:

1. standard diagenetic dewatering by compaction of pore space (consolidation), in a slowly progressive process, mainly caused by continuing sedimentation and subsidence in the basin (Fig. 7);
2. episodic expulsion of brine by increasing overpressure in sand lenses that are enveloped by clay



Fig. 8. Mud eruptions.

beds of low permeability; this can be observed as mud-eruptions (Fig. 8);

3. displacement of deep, hot and therefore relatively light brines by cold, heavy groundwater infiltrating from topographically elevated basin margins; this hypothesis assumes that permeable rocks or active extensional faults provide linked up flow paths (Fig. 9);

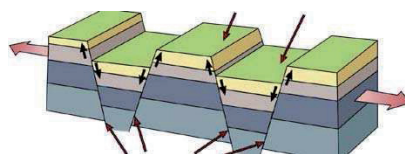


Fig. 9. Extensional faults.

4. tectonic expulsion of connate/formation fluids by a moving nappe pile, causing flow of diagenetic and metamorphic fluids towards the foreland (Fig. 10), where transported heat and dissolved matter result in mineral deposit formation; a present day analogue are methane-rich hot fluids ejected from the accretion wedge at many subduction fronts (Fig. 10);

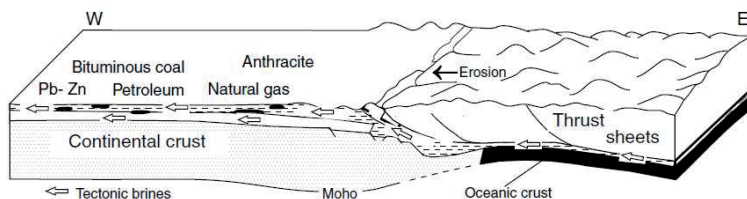


Fig. 10. Flow path of connate/formation water along subduction fronts.

5. fluid overpressures can also be induced by pulses of hydrocarbon generation caused by a rapid thickness increase of cover sediments and consequently of rock temperature; or by a rapid fall of sea levels;
6. expulsion of formation waters by deep heatflow (e.g. cryptic intrusions, underplating).

Many diagenetic mineral deposits display strikingly banded textures of newly-formed precipitates that have been called "diagenetic crystallization rhythmities" (DCR) or "zebra textures" (Fig. 11). The thickness

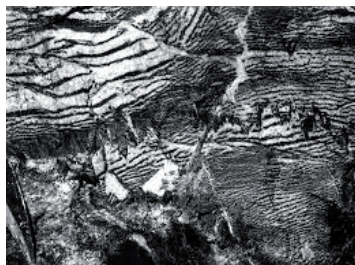


Fig. 11. Zebra textures.

of the bands varies between millimetres and decimetres. They occur in dolomite, evaporite, in sparry magnesite, in siderite and most typically, in stratabound Pb-Zn-Ba-F deposits hosted by platform carbonates.

2- The European Copper Shale

The European Copper Shale is euxinic sediment deposited in a shallow sea of only a few hundred metres water depth (Fig. 12). The European Copper Shale is a black laminated marly claystone formed from clay minerals, dolomite, organic matter, sulphides, anhydrite or gypsum, phosphates and little quartz. It rests on red sand/conglomerate, but in some marginal



Fig. 12. European Copper Shale.

exposures Copper Shale rests directly on folded basement. The European Copper Shale is overlain by carbonates, anhydrite and salt rock. Genetic interpretations of Copper Shale ore formation have always been dominated by the controversy between syn- and epigenetic models.

• Genesis of European Copper Shale

The diagenetic model of Copper Shale ore formation include hot, oxidized (e.g. sulphate-bearing) Ca-Mg-K-Cl connate/formation waters flowing up from deeply buried red bed sediments and volcanics where they acquired their metal contents

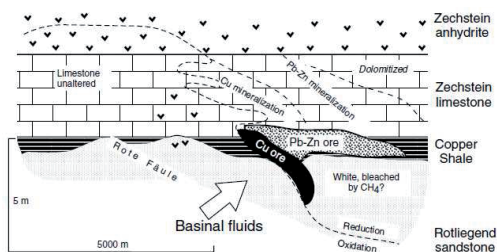


Fig. 13. Genesis of European Copper Shale.

(Fig. 13). Increasing compaction and the low permeability of the hanging evaporites, acting as a seal, focused the flow towards basin margins where the metals were precipitated. Possible agents of ore precipitation from the hot connate/formation waters are:

- i. decreasing T and P;
- ii. mixing with cool, alkaline, meteoric (or sea-) water and
- iii. reduction by contact with organic matter and sulphides of the shale; and methane in the sandstone.

The Copper Shale is dissected by faults/fissures. It is believed that they were the pathways for the hot connate/formation water fluids (Fig. 14). The bulk of the extracted ore, however, is stratiform bodies within the shale containing the ore minerals chalcopyrite, bornite, chalcocite and marcasite. However, some copper orebodies extend from bleached footwall sandstone through shale to hanging wall carbonate. The copper province of White Pine, Michigan, USA and the African Copper Belt of the Democratic Republic Congo and Zambia are comparable but not identical to the European Copper Shale.



Fig. 14. Bedding of European Copper Shale Shale.

3- Diagenetic-hydrothermal carbonate hosted Pb-Zn (F-Ba) deposits (Mississippi Valley Type - MVT)

Lead-zinc ore deposits hosted by marine carbonate rocks are a large and heterogeneous group. They are formed by diagenetic processes. Provinces with more homogeneous characteristics are commonly designated as “types”. Mississippi Valley type (MVT) province, hosts the most intensively investigated lead-zinc deposits (Fig. 15). The paragenesis of diagenetic



Fig. 15. Lead-zinc MVT ore deposits.

hydrothermal carbonate-hosted deposits comprises galena and sphalerite, with a gangue of barite, fluorite, pyrite, marcasite, calcite, dolomite, aragonite, ankerite, siderite, quartz, colloform silica (chert) and occasionally, bitumen.

• Genesis of MVT

MVT Pb-Zn originates diagenetically by mixing hydrothermal diagenetic water with near surface waters that contained reduced sulphur during diagenesis. The diagenetic process was encouraged by bacterial sulphate reduction and consumption of organic matter of the carbonate rocks. Diagenetic ore fluids were derived mainly from evaporated seawater and were driven within platform carbonates by large-scale tectonic events. In addition, magmatic fluids rich in fluorine were mixed into the diagenetic fluids (Fig. 16).

Diagenetic-hydrothermal carbonate hosted Pb-Zn mineralization is closely related to basinal evolution.

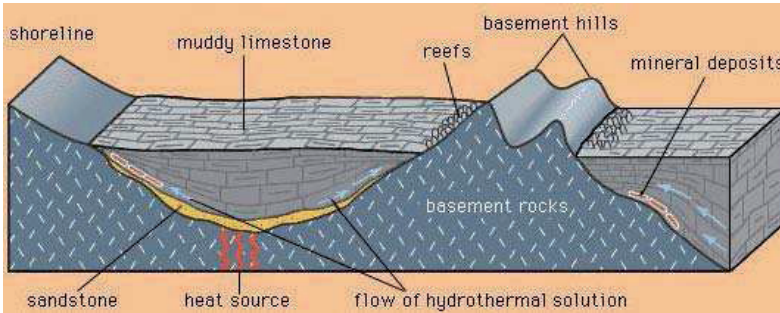


Fig. 16. MVT deposits are always in limestones and are generally located near the edges of sedimentary basins. The hydrothermal solutions that introduced the ore minerals flowed through the sandstones and conglomerates which underlie the limestones. Where they met a barrier to flow, such as a basement high or a basin edge, the solutions moved and reacted with the limestone, depositing ore minerals.

- **Characteristics of MVT deposits**

1. they are epigenetic;
2. they are not associated with igneous activity;
3. they are hosted mainly by dolostone and limestone, in platform carbonate sequences and usually located at flanks of basins rock-dominated passive margin sequences;
4. the dominant gangue minerals are sphalerite, galena, pyrite, marcasite, dolomite, and calcite, whereas barite is typically minor and fluorite is rare;

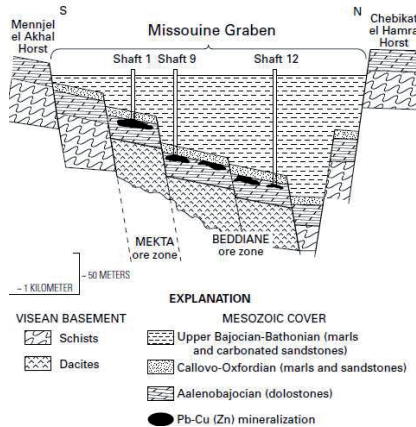


Fig. 17. MVT deposits at Basin flanks.

5. they occur in platform carbonate sequences commonly at flanks of basins or foreland thrust belts (Fig. 17);
6. they are commonly stratabound but may be locally stratiform (Fig 18); and include tectonic or karst cavity breccias that were cemented

by ore and gangue minerals; replacement bodies and cross-cutting veins.

7. the ore fluids were basinal brines with ~10 to 30 % salts; with metal contents reach several thousand ppm.

8. they have crustal sources for metals and sulfur;

9. the most important ore controls are faults and fractures, dissolution collapse breccias, and lithological transitions (Figs. 17 and 19);

10. temperatures of ore deposition are typically 75°C to about 200°C;

11. sulfides are coarsely crystalline to fine grained, massive to disseminated;

12. the sulfides occur mainly as replacement of carbonate rocks and to a lesser extent, open-space fill.



Fig. 18. Stratiform MVT deposits.

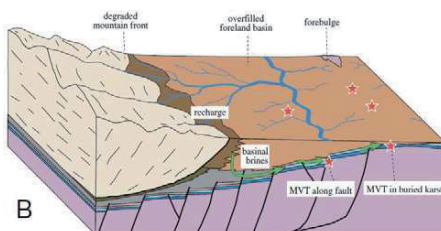


Fig. 19. MVT deposits at Basin flanks.

4-Diagenetic-hydrothermal ore formation related to salt diapirs

Basinal brines and salt rocks, which are assimilated by magmatic intrusions and consequently leach metals from pervaded rocks, may be the key for understanding the iron oxide-copper gold class of deposits. Brines formed by dissolution and/or dehydration of salt rich evaporites within basinal sediments are of specific origin. Most sedimentary basins contain common evaporitic rocks such as anhydrite and rock salt (Fig. 20). Dehydration water from evaporites and formation water in contact

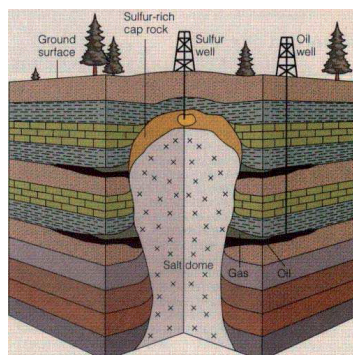


Fig. 20. Salt diapir.

with these rocks must acquire the character of saline brine. Other sources of brines in sedimentary basins include evaporated seawater, terrestrial saline lakes and mature formation water. The source of salinity is discerned by determination of halogen and electrolyte ratios.

- **Genesis of ores related to salt diapirs**

In addition to oil, some salt diapirs are intimately related to ore deposits (Pb, Ag, Zn and Fe in North Africa), and to non-metallic mineral deposits (barite, fluorite, strontianite; emerald in Colombia). Several large metasomatic diapir-related ore deposits represented by siderite and numerous small lead-silver-zinc deposits are found in Early Cretaceous limestones where these are in contact with apical parts of the diapirs in Morocco.

The siderite replaces Aptian limestone that is in direct contact with salt rock (Fig. 21). The origin of these ores is believed to represent a complex interplay of halokinesis (the movement of salt bodies) with fluid generation by diagenesis of evaporites and host rocks (Fig. 21). Partial admixture of meteoric or seawater and brines sourced from the diapir are all possible. At elevated temperature, iron is extremely soluble in reduced, acidic and

highly concentrated brines. Precipitation of siderite is induced by drastic pH-change in contact with the carbonates and by sudden lowering of pressure and temperature. In this model, the diapirs and their solution breccia envelope are preferred channels for upflowing deep diagenetic brines. In active halokinetic phases diapir salt is permeable.

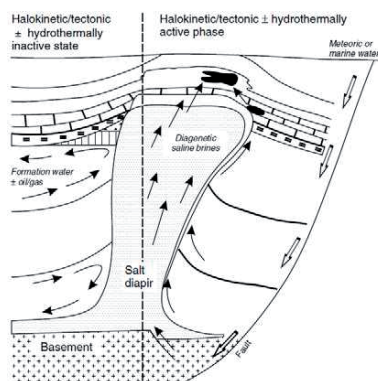


Fig. 21. Diagenetic metallogenesis related to brines released from salt diapirs, based on observations in the Maghreb region in North Africa.

Chapter 7: Metamorphic Ore Deposits

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1- Metamorphic Ore Deposits

Metamorphic ore deposits is a class that owes its economic interest to largely isochemical metamorphic re equilibration and recrystallization of pre-existing material which had no use in its original state. (e.g., transformation of alumina-rich claystones to kyanite/sillimanite deposits, or graphite flakes formed from dispersed bitumen).

2- Contact Metamorphism

Contact metamorphism of ore in the heated zone around magmatic bodies is usually static (i.e., in the absence of dynamic deformation) (Fig. 1). Exposure to high temperatures (with a maximum of 750°C) affects fabric, mineralogy and mineral chemistry (e.g. by driving off water and other volatiles). Fabric changes are confined to a general increase in grain size with rising temperature. Monomineralic ores recrystallize into triple grain boundary junctions at angles of 120 °C. Increase of grain size by metamorphism is important in practice, because processing of coarse ore is less energy intensive.

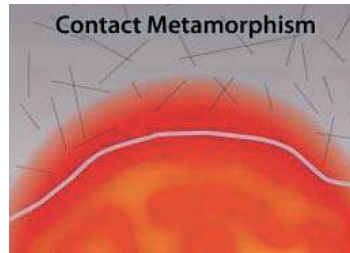


Fig. 1. Contact metamorphism.

During contact metamorphism, the sulphur release from pyrite may induce formation of metamorphic pyrrhotite or even magnetite. Iron oxide ore at contacts may recrystallize to a different oxidation state (e.g. haematite to magnetite) controlled by the oxygen activity imposed by magma or by heated country rocks.

Skarn and contact-metasomatic ore deposits are intimately related to thermal aureoles of magmatic intrusions (Fig. 2). They may be said to be products of contact metamorphism, but the causal agent is the interaction with magmatic fluids and not simple change by heating. Therefore, it was discussed in the magmatic domain.

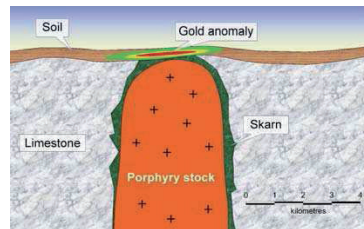


Fig. 2. Skarn.

3-Orogenic (regional) metamorphism

Orogenic (regional) metamorphism of ore deposits is common. Temperatures may reach 1100 °C and pressures 30 kbar. Under these conditions, volatiles (water, etc.) are partly to wholly (at very high metamorphic grades) removed from the system. Metamorphic rocks exhibit grain coarsening, preferred orientation of minerals and a penetrative fabric (e.g. schistosity, foliation) (Fig. 3). Under regional metamorphic conditions, oxides, especially of iron and manganese, react readily with carbonate and silicate minerals. This caused, for example, formation of the diagenetic-metamorphic skarn rocks in Sweden. Elevated Mn-contents of metamorphic silicates (garnet, pyroxene, stilpnomelane, etc.) conserve the geochemical halos of SedEx ore deposits and are useful prospecting tools.

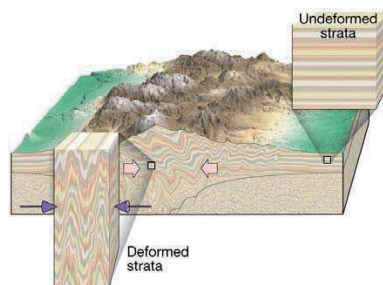


Fig. 3. Regional metamorphism.

Under regional metamorphic conditions, oxides, especially of iron and manganese, react readily with carbonate and silicate minerals. This caused, for example, formation of the diagenetic-metamorphic skarn rocks in Sweden. Elevated Mn-contents of metamorphic silicates (garnet, pyroxene, stilpnomelane, etc.) conserve the geochemical halos of SedEx ore deposits and are useful prospecting tools.

Ores could also be formed from metamorphic fluids. The metamorphic fluids can be considered as solutions that are in equilibrium with host rocks, and although they are dilute, their sheer mass allows significant transfer of dissolved matter. Metamorphic fluids originate primarily by chemical release (devolatilization). Increasing metamorphism, from sub-greenschist facies to anatexis, produces a steady flow of metamorphic dehydration fluids and a decrease of volatiles in the respective metamorphic rocks. Shales contain 4 wt.% water in contrast to mica schist with 2%. By exothermic reactions, basalt assimilates water during the formation of greenschists to a maximum of 13%. Siliceous carbonates lose CO₂ due to generation of metamorphic calcsilicate minerals.

Water in metamorphic rocks occurs mainly in OH-groups of hydrous minerals and in fluid inclusions. Grain boundaries also host tiny inclusions. Some fluids fill open fissures and the pore space. Fluids liberated from rocks undergoing prograde metamorphism acquire economically interesting trace metals (e.g. gold) or other elements (arsenic), either together with the volatiles H₂O, F and Cl from the lattice of transforming minerals.

Chapter 8: Metallogeny and Plate Tectonics

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1- Metallogeny and plate tectonics

The new concept of plate tectonics recognized that the lithosphere is divided into a number of rigid plates, which display considerable lateral movement (Fig. 1). The engine of plate tectonics is convective cooling of the mantle. The resulting lithosphere is in part recycled back into the mantle. Extensional and compressional interactions at plate boundaries are the cause of profusely fertile metallogenetic systems.

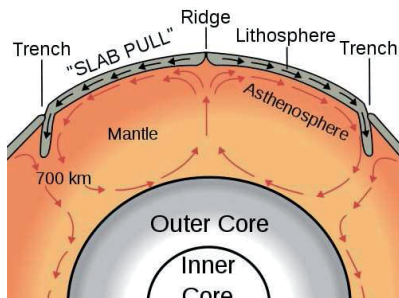


Fig. 1. Lateral movement of plates.

2- Incipient Divergent Plate Boundaries - Extensional Setting (The formation of intracontinental rifts, aulacogens and large sedimentary basins)

Rifts originate by extensional deformation of lithospheric plates and may or may not evolve into a new plate boundary. Rifting causes:

- i) thinning of the crust,
- ii) upflow of hot mantle and
- iii) updoming of rift shoulders.

Volcanic activity within the rifts is often organized into large volcanic centres "hot spots" (Fig. 2). Hot spots can be the origin of three diverging rifts (triple junction). Two of the three rift arms may widen to form a new ocean, whereas the third remains inactive and is called a failed rift arm (Fig. 3). Several failed rift arms display thick sediments. Also, considerable intrusive activity may occur. Settings like this have been called aulacogens (failed arm).

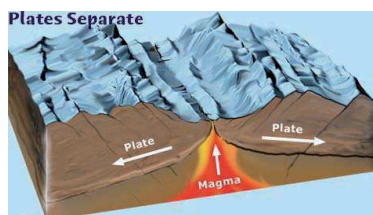


Fig. 2. Beginning of a rift.

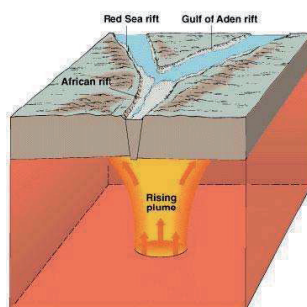


Fig. 3. Triple junction.

If continents rift apart to form ocean basins, other oceans must close. This may be repeated throughout Earth history. Example: the IAPETUS ocean between England & Scotland in the Lower Palaeozoic, closed in the Caledonian; later opening of the Atlantic, almost in the same place.

The cycle is known as the Wilson Cycle (Fig. 4):

1. Rifting of continents by mantle diapirism;
2. Continental drift, seafloor spreading & formation of ocean basins;
3. Progressive closure of ocean basins by subduction of ocean lithosphere;
4. Continental collision and final closure of ocean basin.

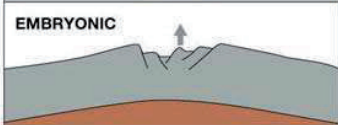
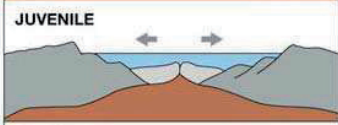
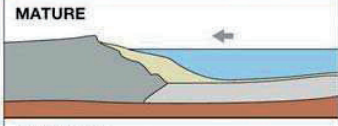
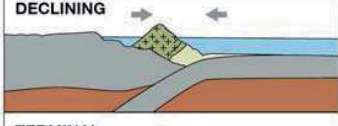
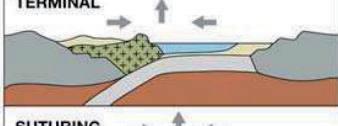
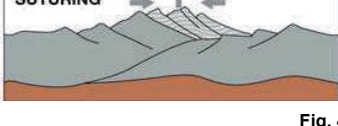
STAGE	MOTION	PHYSIOGRAPHY	EXAMPLE
 EMBRYONIC	Uplift	Complex system of linear rift valleys on continent	East African rift valleys
 JUVENILE	Divergence (spreading)	Narrow seas with matching coasts	Red Sea
 MATURE	Divergence (spreading)	Ocean basin with continental margins	Atlantic and Arctic Oceans
 DECLINING	Convergence (subduction)	Island arcs and trenches around basin edge	Pacific Ocean
 TERMINAL	Convergence (collision) and uplift	Narrow, irregular seas with young mountains	Mediterranean Sea
 SUTURING	Convergence and uplift	Young to mature mountain belts	Himalaya Mountains

Fig. 4. Wilson Cycle.

• **Types of ores in Incipient Divergent Plate Boundaries:**

1. Sediments of continental rifts include terrestrial/alluvial clastic infill that can contain uranium, placers and coal deposits.
2. Freshwater, saline or marine-influenced lake stage succeeds with beds of salt, gypsum, magnesite, phosphate, valuable clays or oil shale.
3. Full marine ingress into the widening rift and inception of oceanic spreading can induce submarine metalliferous exhalation of the black smoker or brine pool type (Red Sea) and the deposition of thick marine sedimentary sequences (Fig. 5).
4. Hot spot-related ore-forming systems include the Bushveld in South Africa, tin-fertile A-granites in Nigeria, many alkali carbonatite igneous complexes with apatite, fluorine, niobium and rare earth element ores (Fig. 5).

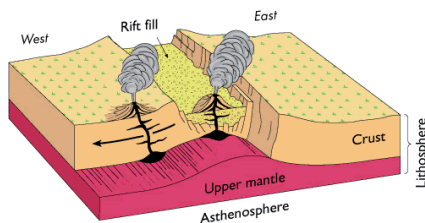


Fig. 5. Extensional tectonics.

5. Rifts can be sites of thick clastic sedimentation. These sediments hold vast amounts of inter-granular salt water (brines). The brines may be in contact with reducing sediments, such as carbonaceous shales, which are also a ready supply of sulphur/sulphate. As the sediments compact, these brines are expelled and can move laterally for large distances until they move up the rift faults. Having been buried deep the brines get hot, and can be very corrosive. So en route they can dissolve considerable amounts of metals. When they rise up the rift faults, the brines cool down and allowing metals to precipitate out. This can be enhanced because oxidising meteoric water (groundwater) may also penetrate down these faults, so metals will be precipitated out when the two meet. Prominent ore provinces include the European Copper Shale which is formed this way (Fig. 6).

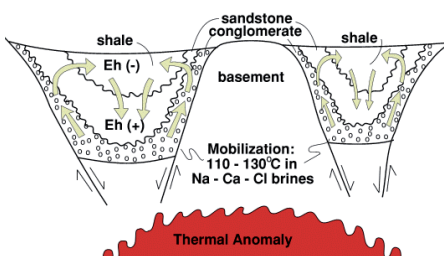


Fig. 6. Copper Shale ore deposit.

6. When rifting reaches the stage of a deep graben with vertical displacement at marginal faults (well into the upper mantle in some cases), hydrothermal convection systems may form. The ascending branch of these hydrothermal systems typically results in deposits of lead, zinc, silver, manganese, fluorine and barite, in the form of veins and metasomatic replacement bodies in rift margin rocks, or of ore beds in the graben sediments. Good examples are many Pb-Zn and Mn occurrences in Tertiary sediments on both sides of the Red Sea.
7. Flood basalt volcanism may be a result of volcanism producing giant Cu-Ni-PGE deposits.
8. Placer gold and Mississippi Valley type (MVT) lead-zinc-barite fluorite deposits are also possible in the rift environment.

3- The evolution of passive continental margins and the disruption of older ore provinces (Divergent Plate Boundaries - Extensional Setting)

The opening of new oceans passes from a high heat-flow rift stage into a marine transgression and thermal contraction phase. Relatively shallow, epicontinental seas may form. As the young ocean widens, passive continental margins develop (Fig. 7). Typical marine epicontinental shelf ore deposits include:

1. salt, phosphate and hydrocarbon source rocks;
2. manganese ore beds;
3. metalliferous marine placers; and
4. banded iron ores of the Superior type.

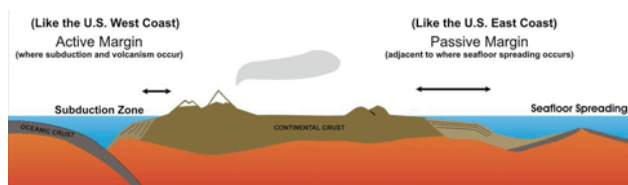


Fig. 7. Active and passive continental margins.

(Epicontinental sea “inland/epiric sea” is a shallow sea that covers central areas of continents, e.g., Caspian Sea – the largest lake on the Earth).

4- Seafloor spreading and the production of new lithosphere at mid-ocean ridges (oceanic divergent, or “Constructive” Plate Boundaries - Extensional Setting)

This concerns ore formation at mid-ocean ridges and in Ophiolites (Fig. 8). After obduction, the products of these processes are ophiolite-hosted deposits. Many ophiolites, however, were not formed at mid-ocean rifts but in tensional settings including back-arc spreading systems, or rifts of primitive island arcs (e.g. the Cyprus ophiolite). Related ores are:

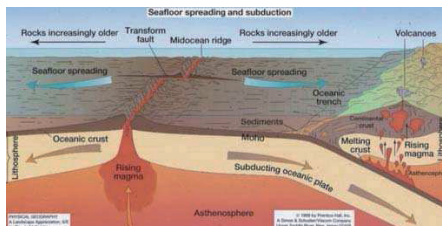


Fig. 8. Active and passive continental margins.

1. sulphide mounds or mud pools;
2. iron-manganese oxides (ochres and nodules); and
3. distal manganese crusts and nodules with important contents of Cu, Ni and Co.

5- Subduction of lithospheric plates at Convergent (“Destructive”) Plate Boundaries (Compressional Settings)

Subduction recycles oceanic lithosphere back into the mantle. The trace of subduction on the seafloor is marked by deep oceanic trenches. Volcanic/Island arcs develop on the overriding plate (Fig. 9).

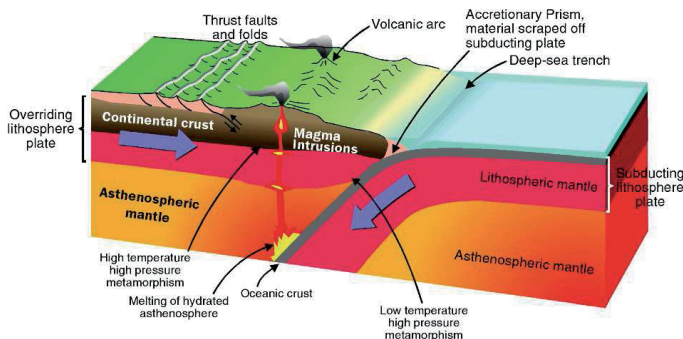


Fig. 9. Subduction of oceanic plate.

It is important to stress that most of the Earth's richest ore provinces are found above subduction zones. There is a great diversity of subduction zone configurations, due to many variables including slab density, thickness and length. Subduction zones show variously:

- i. high or low trench ward plate velocities;
- ii. trench retreat (or more rarely trench advance) velocities;
- iii. slab dip angles and so forth.

There are four main subduction zone configurations:

1. The highly significant Andean type (ocean-continent) collisional margins (Fig. 10) are the sites of the great porphyry Cu–Mo provinces of the world, while inboard of the arc significant Sn–W granitoid-hosted mineralization also occurs. The volcanic regions above the porphyry systems are also the sites of epithermal precious metal mineralization (gold) (Fig. 11).

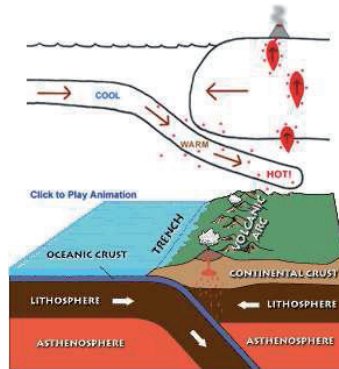


Fig. 10. Ocean-continent collision margin.

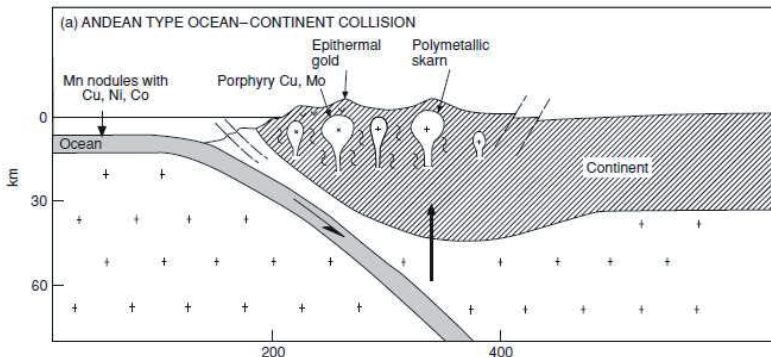


Fig. 11. Mineral ores at ocean-continent collision margin.

2. A similar tectonic setting can exist between two slabs of oceanic crust (ocean-ocean collision) (Fig. 12). In this environment, porphyry Cu–Au deposits occasionally occur associated with the early stages of magmatism in these settings (Fig. 13), whereas the later, more evolved calc–alkaline magmatism gives rise to Kuroko-type VMS deposits. The back-arc basins represent the sites of Besshi type VMS deposition.

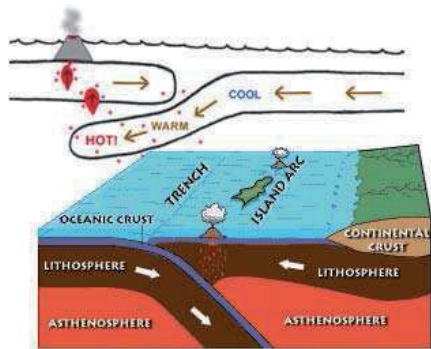


Fig. 12. Mineral ores at ocean-ocean collision margin.

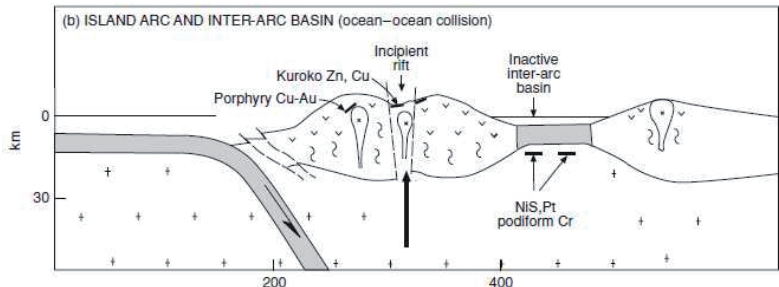


Fig. 13. Mineral ores at ocean-ocean collision margin.

3. Arc–arc collision in the back-arc environment can also result in the preservation of obducted oceanic crust within which podiform Cr and sulfide segregations might be preserved (Fig. 14).

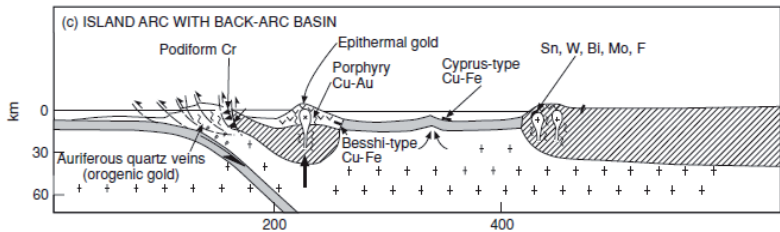


Fig. 14. Mineral ores at arc-arc collision margin.

4. As the arc and continent accrete (become closer), ophiolite obduction can occur, and felsic magmatism may give rise to large-ion lithophile element mineralization (LILE: Ba, K, Rb, Cs, Ca, Sr) (Fig. 15). Ultimately, the oceanic crust is totally consumed to form a zone of continent–continent collision. The process results in thickened crust below collisional belts and the formation of anatectic S-type granitoid melts (Fig. 16). From these melts, Sn–W–U mineralization could occur. In addition, orogeny-driven fluids give rise to orogenic vein-related Au systems (orogenic gold deposits) which could be associated with MVT Pb–Zn deposits.

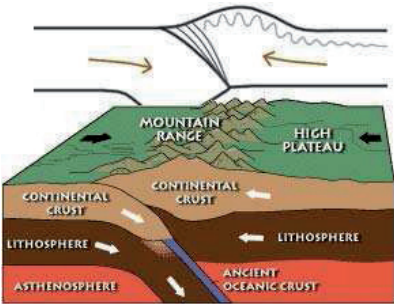


Fig. 15. Mineral ores at arc-arc collision margin.

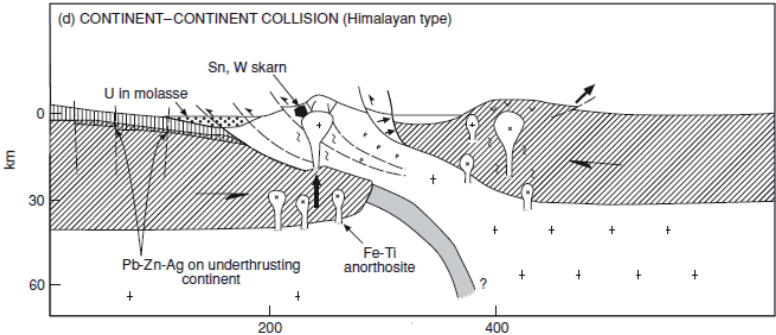


Fig. 16. Mineral ores at continent-continent collision margin.

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